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CONTENTS

	Pages
Stratigraphy and archeology of some Florida soils. By Chas. B. Hunt and Alice P. Hunt	797–806
Devonian-Mississippian transition in central Texas. By P. E. Cloud, Jr., V. E. Barnes, and W. H. Hass	807–816
Submerged river system of Chesapeake Bay. By John T. Hack	817–830
Pleistocene geology of the Door Peninsula, Wisconsin. By F. T. Thwaites and Kenneth Bertrand	831–880
Some pyroxenes associated with pyrometasomatic zinc deposits in Mexico and New Mexico. By Victor T. Allen and Joseph J. Fahey	881–896
Geophysical investigations in the eastern Caribbean: Trinidad shelf, Tobago trough, Barbados ridge, Atlantic Ocean. By J. E. Ewing, C. B. Officer, H. R. Johnson, and R. S. Edwards	897–912
Peneplains and pediplains. By Henri Baulig. (Translated by C. A. Cotton)	914–930
Short Note	
Paleomagnetic survey of Triassic rocks from Arizona. By Paul R. Kintzinger	931–932

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PAPERS IN PRESS FOR FORTHCOMING ISSUES

MARINE GEOLOGY OF THE SOUTHERN HAWAIIAN RIDGE. By E. L. Hamilton

ORIGIN OF PEBBLY MUDSTONES. By J. C. Crowell

UPPER MIDDLE ORDOVICIAN STRATIGRAPHY. By M. P. Weiss

PLEISTOCENE BEDS NEAR THE EDGE OF THE CONTINENTAL SHELF, SOUTHEASTERN LOUISIANA. By W. H. Akers and A. J. J. Holck

GEOLOGIC EVOLUTION OF THE BEARTOOTH MOUNTAINS, MONTANA AND WYOMING. PART 1: ARCHEAN HISTORY OF THE QUAD CREEK AREA. By F. Donald Eckelmann and Arie Poldervaart

CRETACEOUS, PALEOCENE, AND LOWER EOCENE GEOLOGIC HISTORY OF THE NORTHERN MISSISSIPPI EMBAYMENT. By R. G. Stearns

HYDROLOGY OF LAKE BONNEVILLE AND SEDIMENTS AND SOILS OF ITS BASIN. By A. J. Eardley, Vasyl Gvosdetsky, and R. E. Marsell

COLLAPSE FEATURES, TEMPLE MOUNTAIN URANIUM AREA, UTAH. By P. F. Kerr, M. W. Bodine, D. R. Kelley, and W. Scott Keys

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STRATIGRAPHY AND ARCHEOLOGY OF SOME FLORIDA SOILS

BY CHAS. B. HUNT AND ALICE P. HUNT

ABSTRACT

The relationships between some soils and some archeological remains in west-central Florida suggest a sequence of three ages of soil development and two ages of archeological remains.

The oldest soil is represented by a weathered zone on phosphate deposits and antedates the earliest recognized human occupation. Superimposed on this weathered zone are some well-developed but much thinner soils of intermediate age, and at the base of these younger soils are distinctive nonpottery (probably pre-Pottery) archeological remains. The topsoil of these soils contains a different assemblage of stone artifacts and associated pottery. Pottery was first introduced into this region shortly before the beginning of the Christian era.

The youngest soils in the area, those developed on deposits that contain pottery, are thin, and the leached layer in them is only slightly developed.

CONTENTS

TEXT

	Page
Introduction	797
Stratigraphic significance of soils and of fossil or archeological remains	798
Hypothesis	799
Stratigraphy	799
Wavellitic soil	799
Leon soil	801
Archeology of sites in Leon soils	802
Upland sandy soils	803
Archeology of pottery sites on sandy soils	803
Modern soils and archeology of sites related to them	805
Conclusion	806
References cited	806

ILLUSTRATIONS

Figure	Page
1. Index map of west-central Florida	798
2. Diagrammatic section illustrating the wavellitic weathered zone in the phosphate deposits	799
3. Cross section from phosphate district south of Plant City to Hillsborough Bay	800
4. Profile of Leon soil containing artifacts and chert flakes	801
5. Stone tools characteristically associated with Leon soils	802
6. Stone tools characteristically associated with pottery sites	804

INTRODUCTION

Some relationships between certain archeological sites and soil profiles in west-central Florida offer promise of being mutually helpful in revealing additional information about the chronology of the soils and the prehistoric occupations in that region. The relationships were observed in the spring of 1953 as part of the U. S. Geological Survey's investigation of the phosphate deposits in Florida. This report, a by-product of the larger investigation, concerns some relations between the soils, archeology, and geology at about 70 sites in Manatee, Hillsborough, Polk, Pasco, and Hernando counties (Fig. 1).

Most of the localities studied are road cuts. In west-central Florida, abundant chert flakes in the surface wash on road cuts or in roadside ditches commonly indicate the presence of a near-by archeological site, and further search generally leads to finding definite artifacts. In the course of the work, however, it was not possible to search exhaustively or to excavate promising localities. The results therefore are incomplete and are offered as hypotheses rather than conclusions.

Stratigraphic Significance of Soils and of Fossil or Archeological Remains

A deposit, whether of sand, silt, or clay, can be no younger than the soil that has developed on it. By soil is meant the zone, usually layered, that has been altered by weathering at the surface of the deposit. The parent material may be much older than the soil on it, for the soil dates from the time the upper surface of the deposit became exposed to weathering.

On the other hand, a deposit can be no older than the fossils or archeological remains contained in it, but it may be much younger than the fossils or artifacts, for these may have been reworked from older deposits. This generalization neglects intrusive occurrences; burrowing animals, for example, can intrude older deposits, and burials can extend into older deposits, but while these are uncertainties they are not the general condition.

Fossil or archeological remains generally provide a maximum age limit, whereas soils provide a minimum limit. The occurrence to-

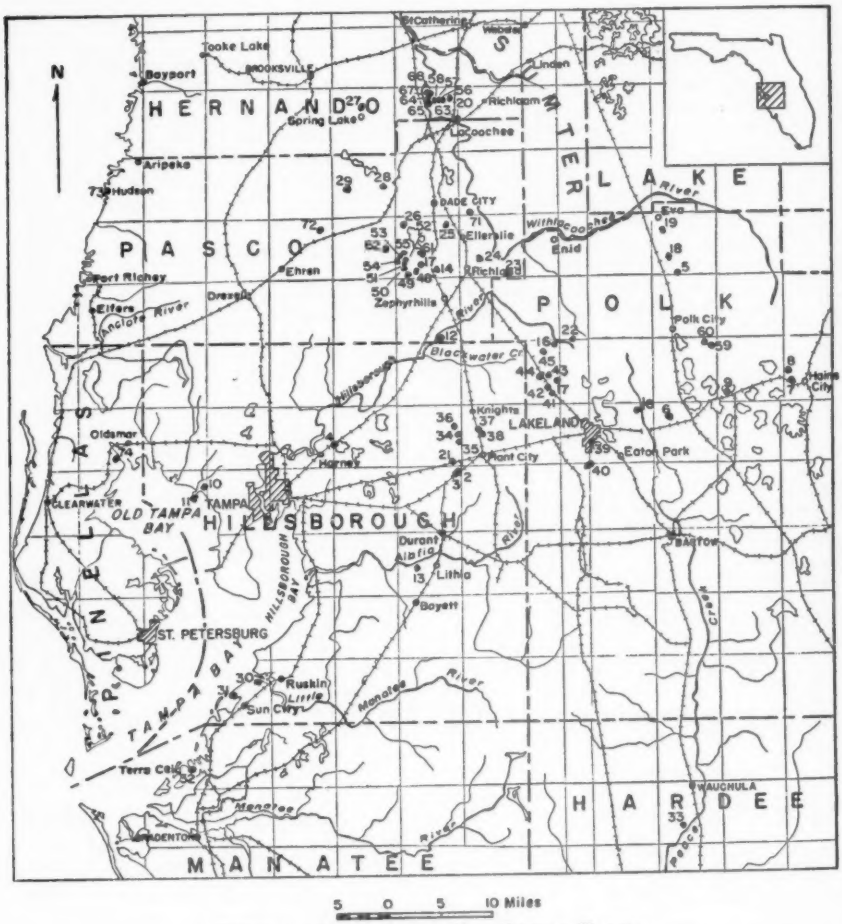

Figure 1.—Index Map of West-Central Florida
Numbers show where archeological remains were found associated with exposures of soil profiles

gether of soils and fossil or archeological remains therefore provides considerable mutual support in developing chronologies; each complements the other.

Hypothesis

Two ages of archeological remains and three ages of soils are suggested by the relationships in west-central Florida.

The oldest soil is represented by the weathered zone on phosphate deposits and antedates the earliest recognized human occupation. Soils of intermediate age, represented by the Leon soil, and some sandy ones like the Lakeland, Norfolk, Blanton, Eustis, and St. Lucie, have developed above distinctive nonpottery archeological remains, but the topsoil of these soils contains a different assemblage of stone artifacts, and these are accompanied by pottery. The youngest soils in the area have developed on deposits containing pottery, which was first introduced into this region shortly before the beginning of the Christian era.

The nonpottery sites, which probably are pre-Pottery in age, were found in two kinds of environment: on poorly drained, swampy ground where the artifacts occur in and below the lower layers of the soils (Leon soil, a ground-water podsol), and on well-drained uplands where similar artifacts occur in or below the lower layers of the sandy soils.

Most of the pottery sites are on the well-drained uplands; at these localities the assemblage of artifacts, which differs from that at the nonpottery sites, is restricted to the upper horizons of the sandy soils. Some of these sites are on stabilized sand dunes. Another group of pottery sites is in the shell mounds along the coast, and these sites provide the clearest examples of the very young soils.

The fact that the three groups of soils are very different, and the fact that the typology of the artifacts varies in an orderly way with respect to position in the soils, strongly indicate that the chronology is as postulated. However, this kind of evidence is limital rather than absolute, and a larger body of data is needed if the interpretation suggested is to be confirmed. For example, the evidence indicates that pottery will not be found beneath a Leon soil (except in an intrusive occurrence), but a large number of observations will have to be made before this kind of negative evidence can be accepted as fact.

Stratigraphy

Wavellitic Soil

The oldest of the soils studied is the deeply weathered, clayey, and wavellitic (aluminum phosphate) zone, 10 feet or more thick, on the phosphate deposits (Fig. 2) (Matson, 1915, p. 21, 65). This weathered zone is developed partly on the Hawthorn Formation of Miocene age (Cooke, 1945, p. 158; Ketner and McGreevey, 1956), partly on the Bone Valley Formation of Pliocene age (Cooke, 1945, p. 203), and partly on deposits that are thought to be early Pleistocene. It is not developed on late Pleistocene deposits; on the contrary, in the coastal belt west of the phosphate district late Pleistocene deposits overlap the wavellitic weathered zone (Fig. 3). The weathering is no younger than late Pleistocene. In part, this weathering may be as old as late Miocene, because in places upper Miocene sand unconformably overlies an aluminum phosphate zone in the Hawthorn Formation (Ketner and McGreevey, 1956).

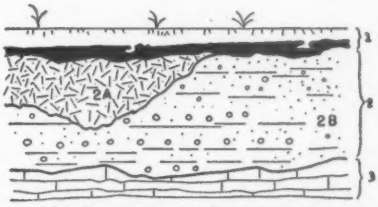

Figure 2.—Diagrammatic Section Illustrating the Wavellitic Weathered Zone in the Phosphate Deposits

Superimposed on the thick zone of wavellitic weathering is Leon soil (1), the profile of which commonly is about 3 feet deep and consists of a thin surface layer over a white leached zone; at the base is a layer of organic matter. The phosphate deposits (2) were developed by weathering of Pleistocene, Pliocene, or older phosphatic formations; the phosphate occurs as wavellite in the upper part of the zone (2*A*) and as apatite in the lower part (2*B*). These zones are measurable in tens of feet. At the base (3) is Miocene phosphatic limestone.

Wavellite is a hydrated phosphate of aluminum apparently derived in this area by weathering of apatite, a fluorine-bearing phosphate of calcium (Altschuler, Cisney, and Barlow, 1952).

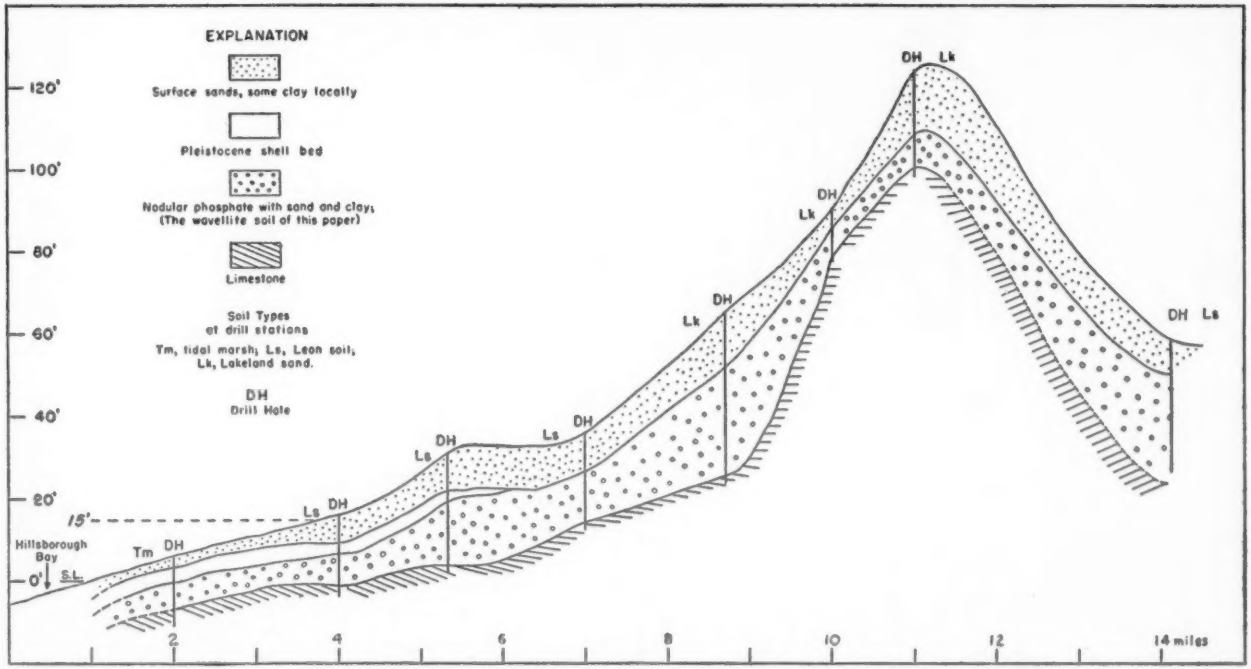

FIGURE 3.—CROSS SECTION FROM PHOSPHATE DISTRICT SOUTH OF PLANT CITY TO HILLSBOROUGH BAY

Based on drill data. The Leon soil and the upland sandy soils are developed on a mantle of sand that, near Hillsborough Bay, overlies a Pleistocene shell bed. This shell bed overlies a deposit of nodular phosphate, the wavellitic zone of this paper. Cross section by R. G. Petersen.

The development of wavellite from phosphatic parent materials may be analogous to the development of laterite from iron-bearing parent materials. Both processes involve the breakdown of clay minerals and loss of SiO_2. We are using the term lateritization to mean "the leaching out and elimination of SiO_2, alkali, and alkaline earths, and the concentration in their hydrated form, of iron and aluminum oxide" (Mohr and Van Baren, 1954, p. 353). Probably the wavellitic weathered zone developed at various times during the Tertiary and during one or another of the interglacial stages of the Pleistocene when lateritic weathering extended far north of where it does today.

The wavellitic soil is older than either the Leon soil or the sandy upland soils, for these are superimposed on the upper horizons of the wavellitic soil and occur on the late Pleistocene deposits along the coast.

Leon Soil

The Leon soil occurs on the poorly drained lands in the region. These lands, flat or gently undulating, are in large part lowlands in or at the edges of the swamps, but they include some poorly drained upland flats. The soil is a ground water podsol—that is, it has an organic layer in the subsoil at the top of the water table (Fig. 4). At the surface there is an ash-gray sandy layer 3–6 inches thick containing finely divided organic matter. Below this is the leached zone, which is light-gray, locally almost white, incoherent sand 1 or 2 feet thick. Below this is the organic layer, 3–9 inches thick, a hardpan containing organic and iron salts (Fowler and others, 1927, p. 28). All layers in the soil are strongly acid. The native vegetation is longleaf pine, slash pine, and saw palmetto.

At several localities many stone artifacts were found in the buried organic layer. The artifacts occur in a way that seems to preclude their being intrusive into the soil (Fig. 4), because no disturbance was detected in the boundary between the organic layer and overlying leached layer. Disturbance of this boundary could be masked by later deposition of organic matter at the top of the hardpan, but this seems improbable because of the wide extent of the artifacts at some localities. At two localities, for example (Fig. 1, localities 45 and 48), the exposures are along the drainage ditches on each side of a road, and the artifacts are distributed for several hundred feet along the outcrop of the organic layer in the ditches.

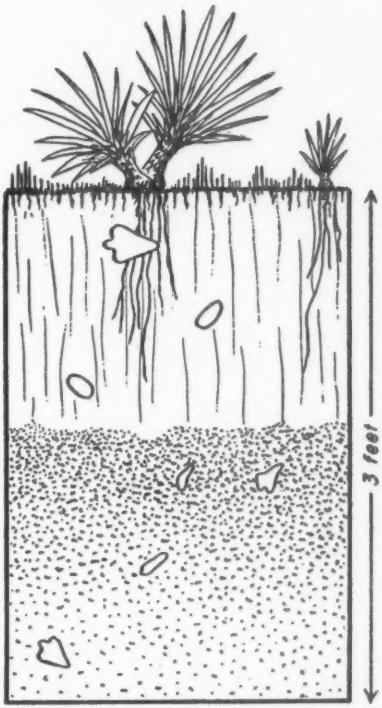

FIGURE 4.—PROFILE OF LEON SOIL CONTAINING ARTIFACTS AND CHERT FLAKES

At the top is an ash-gray sandy layer 1 or 2 inches thick; below this is the light-gray or almost white leached zone, commonly about 18 inches thick. In the subsoil, 20–30 inches below the surface, is a dark layer of organic matter. At localities 45 and 48 in Figure 1 flakes and artifacts were found at all levels of the soil; probably the artifacts antedate the soil, because some were found deep in the apparently undisturbed organic layer.

Artifacts that are similar typologically and made of similar materials occur also in the upper horizons of the Leon soil, but their position does not indicate that they are older than the soil. However, the common occurrence, in extensive areas, of specimens deep in the buried organic layer does suggest that the

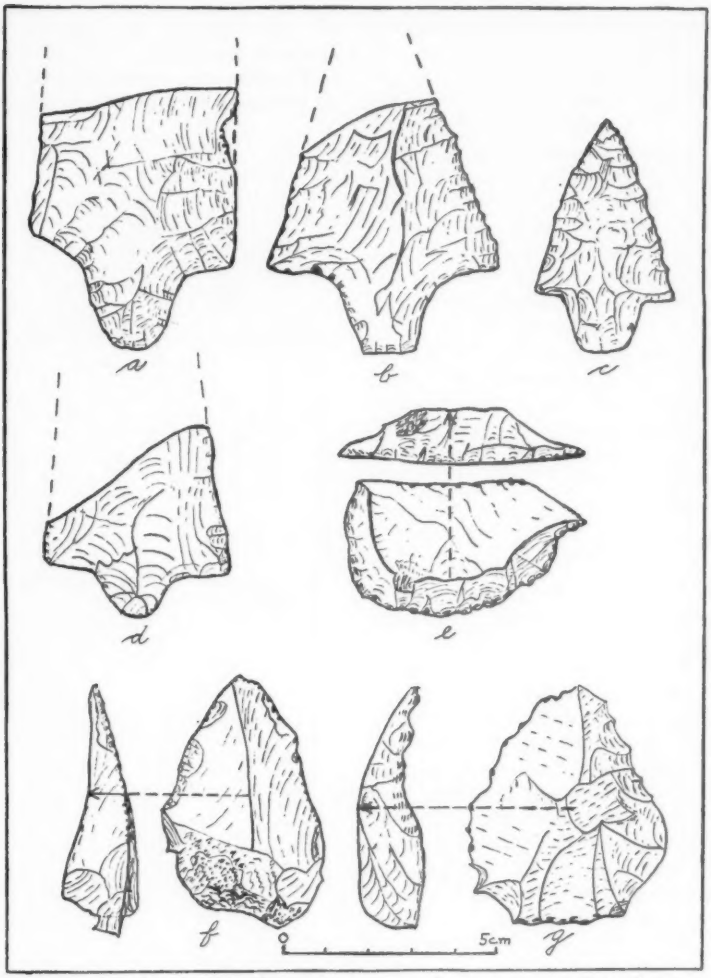

FIGURE 5.—STONE TOOLS CHARACTERISTICALLY ASSOCIATED WITH LEON SOILS

soil at those places is younger than the occupation represented by these artifacts.

Archeology of Sites in the Leon Soils

Many stone flakes and cores and some definitely shaped stone artifacts (Fig. 5) were found in the deep layers of the Leon soil at a dozen localities. Three (Fig. 1, localities 45, 48, and 50) yielded abundant flakes, cores, and definitely shaped artifacts from all levels of the soil including the hardpan layer. Eight other localities (Fig. 1, 16, 49, 51, 52, 53, 54, 55, and 72) yielded less definite stratigraphic information but probably are similar to the others; some (Fig. 1, localities 16, 51, 52, and 55) are only surface sites.

Sites 45 and 48 are on broad, low, poorly drained mounds about 1000 feet in diameter. Each is surrounded by swamp. Judging from the amount and kind of stonework they were occupied for a long time as camp sites. Sites 49

to 55 may have been quarries, for near by there are outcrops of the kind of chert that was favored for making the artifacts.

Most of the stone tools from these sites are amorphous and crudely chipped. Figure 5 shows the most common types. Projectile points are heavy and large, with contracting stems and wide, straight, or sloping barbs. They must have been used with spear or atlatl, as they are too large to have been used with a bow. Some may be stemmed knives. Scraping tools and picks (Fig. 5, *f*, *g*) are common, and a number are combination graving and scraping tools (Fig. 5, *e*). In this tool, besides the graving point, one side has a wide-angle scraping or cutting edge. Other tools found but not illustrated are unworked flake knives and oval knives, showing nicks from use, and chopping tools.

Artifacts like some of those found with the Leon soils occur also in sites of Pottery age in Florida. Similar projectile points have been found with pottery representative of the periods referred to as Perico Island, Santa Rosa-Swift Creek, Weeden Island, and Safety Harbor (Willey, 1949, Pl. 16, *j*; Pl. 24, *d*; Pl. 42, *f*; Pl. 55, *e*). However, the absence of pottery at the sites associated with the Leon soils, the absence of the later type projectile points and other late type stone tools (Fig. 6), and the crudity and lack of variety of the tools found support the geologic evidence—the occurrence of the artifacts in the subsoil—that these sites in the Leon soil are pre-Pottery.

Upland Sandy Soils

Loose sandy soils are extensive on the well-drained, rolling uplands, and similar soils occur on some stabilized sand dunes west of the Withlacoochee River in Pasco County. The sandy upland soils are referred to as the Lakeland, Norfolk, Blanton, Eustis, and St. Lucie sand, or fine sand (Fowler and others, 1927). They grade laterally into the Leon soil (Fig. 3) and are believed to be contemporaneous with it.

These soils consist of a 2- or 3-inch surface layer that is dark gray, grayish brown, or yellowish gray, probably because of finely divided organic matter. In the fine sand this layer may be loamy. In the Norfolk sand the surface layer grades downward rather abruptly into a layer of pale-yellow sand 1 or 2 feet thick, and this in turn grades downward into yellow sand 2 or 3 feet thick. The Blanton and St. Lucie are lighter-colored; the Eustis is browner. Except in the areas of stabilized dunes the subsoils generally contain some clay. The soils are strongly acid. The native vegetation is pine and oak.

Archeological remains recording a Pottery age occupation were found in the surface layers of these soils at 28 localities. There is little doubt that these soils are older than the pottery occupations.

At a few localities artifacts like those associated with Leon soils were found deep in the subsoil. At site 25 an unfinished projectile point like *d* in Figure 5 was found 3 feet below the surface. However, the evidence that these sandy soils are younger than the deeply buried artifacts is incomplete, although there are enough occurrences similar to those in the Leon soils to suggest that these well-developed sandy soils are at least roughly correlative with the Leon.

Archeology of Pottery Sites on Sandy Soils

Pottery and an assemblage of stone artifacts different from those in the Leon soil were found in the upper layer of the sandy soils at many localities. Eleven of these sites are on uplands (Fig. 1, localities 5, 6, 7, 22, 24, 28, 38, 42, 44, 62, 83). Seven others are on stabilized dunes (Fig. 1, localities 20, 27, 57, 64, 67, 68, 71). A considerable number of other sites yielded similar stonework but no pottery (Fig. 1, localities 3, 4, 8, 9, 14, 17, 18, 19, 21, 23, 26, 29, 34, 35, 36, 37, 43, 46, 47, 56, 58, 60, 63, 65, 66). The pottery sites evidently postdate the sandy soils.

Most of the sites are less than 100 feet in diameter and have a few stone tools, flakes, and sherds of pottery on them. Evidently most were used only briefly as camp sites. Some, however, like 17 and 62, are extensive and may have been small villages.

Pottery containing mineral temper (as distinct from fiber temper) is thought to have been introduced into this part of Florida a short time before the beginning of the Christian era (Bullen, 1952, p. 5). Fiber-tempered pottery was

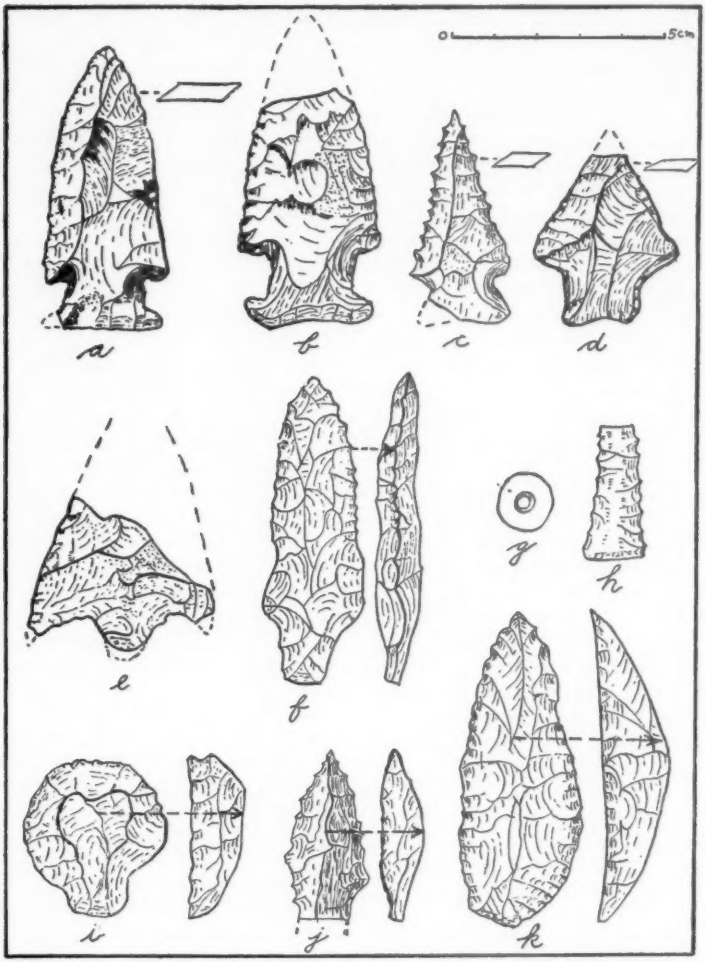

FIGURE 6.—STONE TOOLS CHARACTERISTICALLY ASSOCIATED WITH POTTERY SITES

introduced earlier, perhaps much earlier, but there is no basis yet for estimating when it was first made. Some of the sites referred to herein as pre-Pottery may have once contained fiber-tempered pottery.

Most of the pottery sherds found are tempered with fine to medium sand. The paste commonly is pitted, as if it once contained fragments of shells or crushed limestone that have been leached out. (*Cf.* Perico Island and Pasco pottery series.) The clay of the paste commonly contains sponge spicules. The sherds examined in thin section evidently were indigenous wares, for they are made of local materials: the sand contains considerable chert which is characteristic of sands in this area and lacks the abundant heavy minerals that characterize sands and sherds from the northwest coast of Florida. The most common sherds are plain gray, brown, or buff. Some are check-stamped (*cf.* Wakulla check-stamped), and some are incised.

The assemblage of stone artifacts from these pottery sites (Fig. 6) differs from that associated

with the Leon soils. Although the projectile points are smaller than those at the sites in the Leon soil, they are long enough and thick enough to be more suitable as spear than as arrow points. Apparently the spear or atlatl was used until very recently in this part of Florida.

Several of the projectile points are of the "spinner" type; they are obliquely beveled on opposite sides of the opposite faces (Fig. 6, *a*, *c*, *d*). Similar opposite-beveled points are reported from central Georgia in the early Swift Creek period which followed the Stallings Island period of fiber-tempered pottery (Fairbanks, 1952, p. 287; Fig. 157, D). Large, side-notched projectile points like *a* and *b* of Figure 6 and a stemmed knife (Fig. 6, *f*) are reported from sites of the historic and protohistoric Safety Harbor period (Willey, Pl. 55, *a*, *d*).

In these sites the variety of tools is greater than in the sites associated with Leon soils, and the tools are more finely flaked. No grinding stones and no European tradeware were found on any sites.

The materials most commonly used for making the stone tools at the pottery sites differ from the materials most commonly used at the Leon soil sites. At the pottery sites the chert most commonly used is flesh-colored, fine-grained, vitreous, microcrystalline chalcedony. The individual crystals commonly range from 1 or 2 microns in diameter to about 20 microns, and they are distributed in patterns that suggest replacement of Foraminifera or mollusk shells. Some of the chalcedony is speckled with black grains, possibly pyrite or iron oxide. Vesicles are common; many are lined with fibrous chalcedony or quartz. Some are lined with hematite. Much of this chert is believed to be from the Ocala limestone of Eocene age, which crops out north of the area covered in this paper.

The chert most commonly used for shaping the artifacts at sites in the Leon soils, however, is dark vitreous chert that evidently replaced coralline limestone, presumably the Tampa limestone of Miocene age which crops out at many places in the area covered by Figure 1. This chert is in part microcrystalline chalcedony with a grain size less than 5 microns and in part chalcedony and quartz in grains up to 75 microns in diameter. The structure of the replaced corals can be seen in most thin sections, and bedding can be seen in some.

Both kinds of chert have developed a distinct patina. The patina on flakes of chert that replaces coralline limestone commonly is 10 mm thick. On the cherts from the Ocala limestone the patina rarely is as thick as 5 mm. The difference in thickness of patina fits with the supposition that one type is more severely weathered and older than the other, but this evidence is of doubtful value because the parent materials are not alike.

Modern Soils and Archeology of Sites Related to Them

The youngest soils in the region consist merely of a humus-enriched layer over a parent material whose upper part is leached so slightly that calcium carbonate shells that were originally present are still intact. Such soils occur on the pottery-bearing shell mounds along the coast, and there are similar ones along the Atlantic seaboard as far north as New England. Similar post-Pottery soils containing fragments of delicate shells of fresh-water or terrestrial mollusks occur on the youngest alluvial deposits along many rivers of the seaboard states. In all these soils in which the leached layer is absent or only slightly developed there is archeological evidence (pottery) and geological evidence (physiographic position) indicating that the period of weathering is measurable in hundreds of years. The evidence suggests that few soils along the Atlantic seaboard or in west-central Florida that preserve abundant shell fragments in the leached zone in otherwise nonlimy parent material antedate the Christian era.

The shell and sand-mound sites along the coast that are mantled by these shallow, feebly developed modern soils have been described by Bullen (1951; 1952), Griffin and Bullen (1950), Willey (1949), and Griffin (1952). The earliest of these sites in west-central Florida (ca. 100 B.C. to 700 A.D.) is believed to be at Perico Island in the southern part of Tampa Bay. Weeden Island, north of St. Petersburg, seems to have been occupied from about 700 to 1400 A.D. At Safety Harbor (Fig. 1, locality 74) the occupation extended into historic time.

The early mounds, whether for burial or habitation, were made of shell. Later, sand was used for the burial mounds, although the villages continued on the shell middens. However, at none of these sites is there a thick or strongly developed leached zone in the soil.

Conclusion

If the stratigraphy of the soils and archeological remains has been interpreted correctly, the history of the area has been somewhat as follows:

At some stage during late Tertiary and early Pleistocene time the land surface of west-central Florida had a configuration like that of the present time, and the climate was such that it produced wavellitic (lateritic?) weathering of the phosphatic limestone formations. The top of this weathered zone extends to sea level and possibly below it (Fig. 3), and sea level was therefore no higher at that time than it is today.

At some later time the area became occupied by a people who fashioned the large crude stone implements illustrated in Figure 5. These people had not learned how to make pottery—or, at least, mineral-tempered pottery. At many places, perhaps generally, this occupation antedates the Leon ground-water podsol. Even if similar soils had formed before this occupation, the fact that in certain places soils having a typical Leon profile have developed since that time indicates that the rate of soil development in that environment is high enough to allow the soil to reach equilibrium in a short time, presumably within the Recent period.

This early occupation probably was at a time when the water table was lower than it is today, for the artifacts extend below the hardpan layer of the ground-water podsol. One can only speculate whether this was during a dry period such as that which occurred when the now stabilized dunes were formed in the north part of the area (early Recent?), or during a glacial period when sea level was lower.

The weathering that produced the Leon soil and the sandy upland soils followed this occupation, and after these soils had become well developed the area was occupied by people who knew how to make mineral-tempered pottery and fashioned stone tools that are more varied and finely flaked than those of the earlier occupation. Along the coast these people gathered quantities of shells and disposed of them in mounds. The slight leaching in the upper surface of these shell heaps, in contrast to the considerable leaching in the Leon soils, suggests a considerable antiquity for the Leon and related upland sites.

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DEVONIAN-MISSISSIPPIAN TRANSITION IN CENTRAL TEXAS

BY P. E. CLOUD, JR., V. E. BARNES, AND W. H. HASS

ABSTRACT

The Devonian-Mississippian transition outcrops of central Texas are here described summarily and assigned to a new stratigraphic unit, the Houy Formation. The beds included are mainly Upper Devonian, but partly Lower Mississippian. Locally a basal fraction may be Middle Devonian. Although the deposits included are diverse and their associations complex, the maximum surface thickness so far known is only about 17 feet.

The principal subdivisions, in their usual ascending order, are the Ives Breccia Member (Plummer *in* Bullard and Plummer, 1939), the Doublehorn Shale Member (new), and a thin unnamed phosphoritic interval. Commonly, however, one or more of these members is absent, and rocks not assigned to any member are present. The Ives Breccia Member includes lag deposits of detrital chert of varied age and source. The Doublehorn Shale Member includes black, fissile, spore-bearing shale of Late Devonian age which in a few exposures grades upward into shale of Early Mississippian age. The phosphatic beds are partly Late Devonian and partly Early Mississippian. Remnants of the Doublehorn Shale Member have been found only along the eastern side of the Llano region, but the other units are more widely distributed, and rocks assignable to the Houy Formation are to be looked for between Ordovician and Upper Mississippian deposits anywhere around the Llano region.

Although most abundant in the upper beds, phosphatic inclusions occur locally throughout the Houy Formation. This gives a stamp of unity to the sequence and distinguishes it from the immediately underlying beds as well as from the overlying Chappel Limestone of later Early Mississippian age. The Houy is also a unit of more than average radioactivity and is readily detected in the subsurface by radioactive drill-hole logging.

It correlates with the Late Devonian and earliest Mississippian black-shale sequences of other Midcontinent and Midwestern areas. Four of the six conodont zones (Hass, 1947; 1956a; 1956b) in deposits of this age in Ohio, Tennessee, Oklahoma, and Arkansas are recognized also in central Texas.

CONTENTS

TEXT

	Page
Introduction	808
General features of the Houy Formation	810
Conodont zones and correlation	812
Reference sections	813
General statement	813
Burnam Branch	815
Houy ranch	815
References cited	815

ILLUSTRATIONS

Figure	Page
1. Geologic maps along Doublehorn Creek, Burnet County, Texas	814

Plate	Facing page
1. Key sections of the Houy Formation	807
	Following page
2. Houy Formation at Burnam Branch and Houy ranch, Ives Breccia Member at type section, and King Spring section	808
3. Houy Formation, Chappel Limestone, and Barnett Formation at Fowler ranch and King Spring section	808
4. Representative conodonts of the Houy Formation	808
5. Representative conodonts of the Houy Formation	808

Introduction

The stratigraphic record of the Llano region, Texas, between Early Ordovician and Pennsylvanian time is a complex of formerly extensive units, the remnants of which have been described piecemeal as they have been discovered. In the case of the transition beds here under discussion, it was the writers' original intention to reserve publication until a comprehensive report on the stratigraphy and paleontology of all these relict Paleozoic formations could be prepared. As completion of such a work still faces considerable delay, it now seems advisable to publish essential data on the Devonian-Mississippian transition beds and bring their nomenclature up to date.

Undescribed or unnamed rocks recently found in the Llano region include, in upward succession, (1) coarse-grained limestone with fossils indicating equivalence to the Onondaga Limestone of Early or Middle Devonian age, (2) additional deposits of Late Devonian and Early Mississippian age, of which a few have been mentioned in print (Barnes, 1951, p. 7; 1956, p. 21–22); and (3) a phosphorite deposit of Early Pennsylvanian age which is the subject of a separate report by Barnes (1954).

Although the megafossils from the coarse-grained Devonian limestone suggest to Cloud a

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Plate 2.—HOUY FORMATION AT BURNAM BRANCH AND HOUY RANCH AND IVES BRECCIA MEMBER AT TYPE SECTION AND AT KING SPRING SECTION

Figure 1.—General view of type section of Houy Formation at juncture of Burnam Branch and Doublehorn Creek, Burnam ranch (Pl. 1, section 8 in part).

Figure 2.—Detail at upper part of section shown in Figure 1 of this plate, at contact between the Houy and Barnett formations. Chappel Limestone is missing and upper hammer rests on thin phosphorite bed at top of Houy Formation (USGS coll. 15552-C). Hole below and to right of lower hammer is site of USGS coll. 3926-SD.

Figure 3.—Greenish silty shale of probable late Middle or early Late Devonian age, dug through at depth of 2 feet, beneath 3 feet of Ives Breccia Member and resting on similar Ives-like chert in Doublehorn Creek on Rubin Houy ranch (Pl. 1, section 9). Ives is early Late Devonian age at this place, and both it and green shale are included in the Houy Formation.

Figure 4.—Blocks of Ives Breccia Member at type site Plate 1, section 2, from a kodaslide by W. Charles Bell.

Figure 5.—Smooth surface of Ives Breccia Member (×3/5) showing large amount of relatively fine matrix, King Spring (Pl. 1, section 3; also Pl. 3, Fig. 6).

Plate 3.—HOUY FORMATION, CHAPPEL LIMESTONE, AND BARNETT FORMATION AT FOWLER RANCH AND KING SPRING SECTION

Figure 1.—Hammer on upper phosphorite of Houy Formation in middle of photograph. Doublehorn Shale Member below (4 feet thick to creek level) and petroliferous shale of Barnett Formation above. Loc. 27T-6-42D, Plate 1, section 7 in part. Chappel Limestone is missing at this site.

Figure 2.—Chappel Limestone 3 feet thick, preserved in small collapse structure 100 feet east of Figure 1 of this plate.

Figure 3.—Trench section through 8 feet of weathered Doublehorn Shale Member of Houy Formation, Ives Breccia Member or highly cherty beds of the Stribling Formation (x) beneath and Chappel Limestone (Cc) above. C_1, C_2, and C_3 indicate subdivisions of loc. 27T-6-42C, equivalent to USGS colls. 3924-SD, 3922-SD, and 15550-C of Plate 1, section 7. Exposure is at left side of Figure 2 of this plate.

Figure 4.—Loc. 27T-6-42B, 500 feet eastward down the draw from Figure 1 of this plate. Four- to 12-inch-thick Chappel Limestone (Cc) arches over Doublehorn Shale Member of Houy Formation from hammer at left to Cc at right.

Figure 5.—Doublehorn Shale Member dug and blasted out beneath 5-inch-thick Chappel Limestone (Cc) near right side of Figure 4 of this plate. B_2 and B_4 indicate subdivisions of loc. 27T-6-42B, corresponding to USGS colls. 3923-SD and 15551-C of Plate 1, section 7.

Figure 6.—King Spring section (Pl. 1, section 3). Hammer at left rests against 2- to 11-inch-thick argillaceous limestone of early Kinderhook age (USGS colls. 15541-C, 15542-C, 15543-C), on top of which (i) is a 12- to 18-inch-thickness of the Ives Breccia Member (USGS coll. 3919-SD), both of the Houy Formation (*see* Pl. 2, fig. 5).

FIGURE 1

FIGURE 3

FIGURE 2

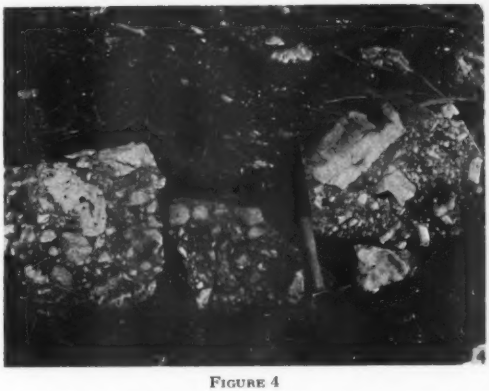
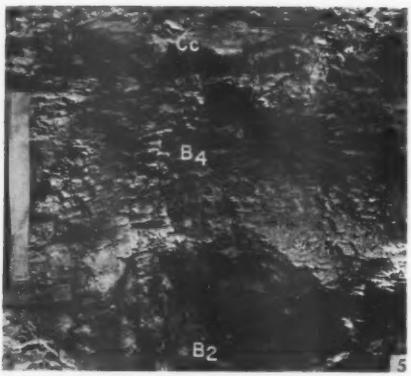


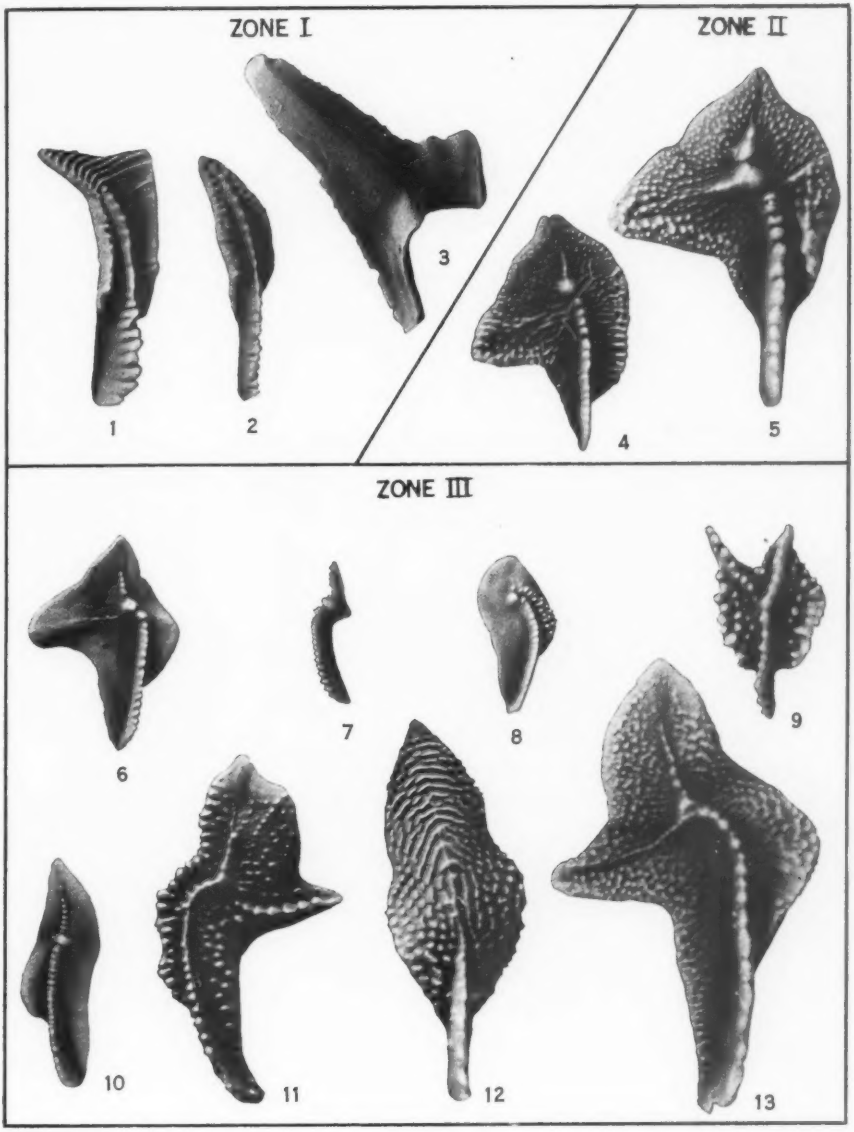
FIGURE 4

FIGURE 5

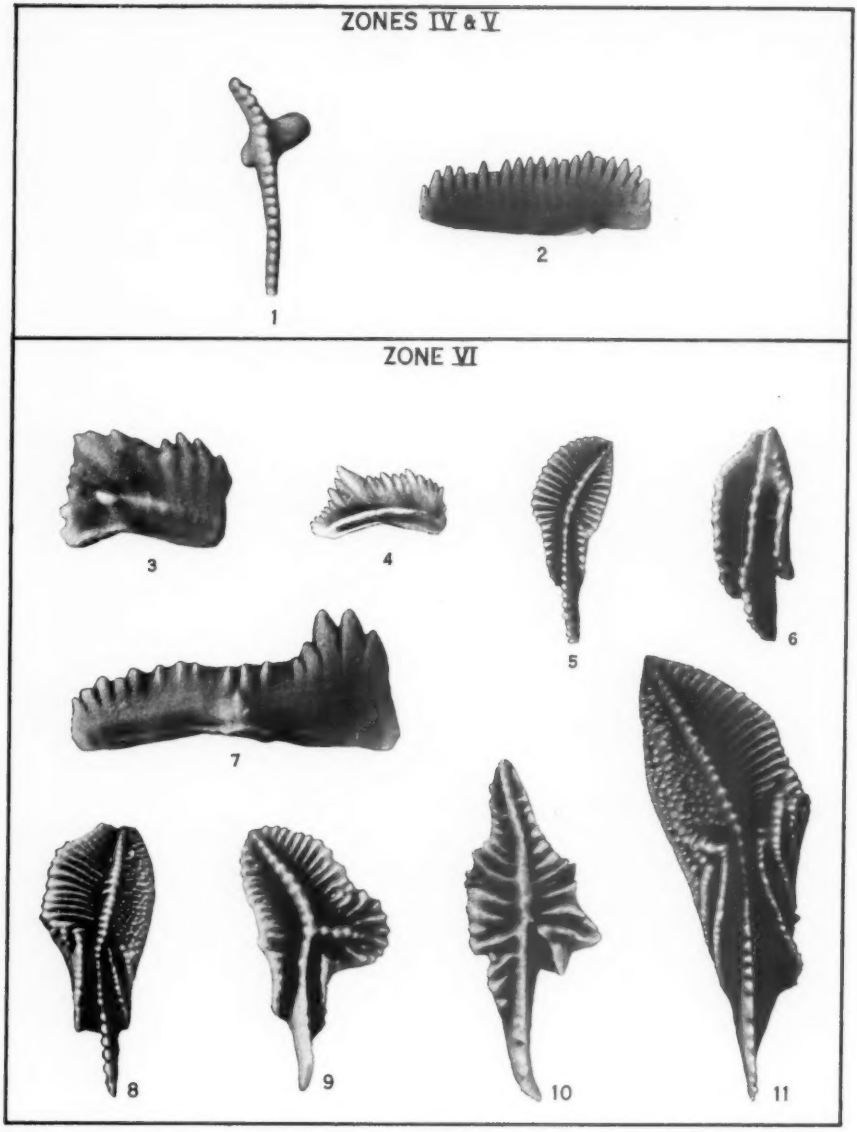
HOUY FORMATION AT BURNAM BRANCH AND HOUY RANCH AND IVES BRECCIA MEMBER AT TYPE SECTION AND AT KING SPRING SECTION



HOUY FORMATION, CHAPPEL LIMESTONE AND BARNETT FORMATION AT FOWLER RANCH AND KING, SPRING SECTION



REPRESENTATIVE CONODONTS OF THE HOUY FORMATION



REPRESENTATIVE CONODONTS OF THE HOUY FORMATION

slightly higher position in the Onondaga sequence than the type Stribling, the field relations so far observed are obscure, and the limestone could be below or equivalent to the microgranular limestone, dolomite, and chert of the Stribling Formation. A point favoring a relatively high position for the coarse limestone is that nothing similar to it has been found between typical Stribling and a thin and irregular interval of sparingly glauconitic impure limestone (Barnes *et al.*, 1953, Fig. 1) that is now provisionally considered basal Stribling.

The phosphorite of Pennsylvanian age, provisionally referred to the basal part of the

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PLATE 4.—REPRESENTATIVE CONODONTS OF THE HOUY FORMATION

All magnifications ×26. All specimens except Figures 2 and 7 from USGS coll. 9264-C (Pl. 1, section 10). Illustrated specimens selected for quality, despite fact that all except those represented by Figures 4 and 5 are considered to have been reworked.

FIGURE 1.—*Polygnathus linguiformis* Hinde. Oral view, hypotype, U.S.N.M. 114904.

FIGURE 2.—*Polygnathus pennata* Hinde. Oral view, hypotype, U.S.N.M. 127851. Locality about 5 miles southeast of Lampasas, in Burnet County, on north side of Pillar Bluff Creek (Barnes *et al.*, 1947, Fig. 3, loc. TF-294), USGS coll. 3940-SD.

FIGURE 3.—*Neoprioniodus alatus* Hinde. Lateral view of inner side, hypotype, U.S.N.M. 127852.

FIGURE 4.—*Palmatolepis unicornis* Miller and Youngquist. Oral view, hypotype, U.S.N.M. 127853.

FIGURE 5.—*Palmatolepis subrecta* Miller and Youngquist. Oral view, hypotype, U.S.N.M. 127854.

FIGURE 6.—*Palmatolepis subperlobata* Branson and Mehl. Oral view, hypotype, U.S.N.M. 114826.

FIGURE 7.—*Palmatolepis gracilis* Branson and Mehl. Oral view, hypotype, U.S.N.M. 114827. Locality about 4 miles southeast of bridge over Colorado River at Marble Falls, Burnet County, along north side of ranch road. USGS coll. 9053-C.

FIGURE 8.—*Palmatolepis quadrantinodosa* Branson and Mehl. Oral view, hypotype, U.S.N.M. 114821.

FIGURE 9.—*Ancyrognathus bifurcata* (Ulrich and Bassler). Oral view, hypotype, U.S.N.M. 114834.

FIGURE 10.—*Palmatolepis glabra* Ulrich and Bassler. Oral view, hypotype, U.S.N.M. 114822.

FIGURE 11.—*Palmatolepis rugosa* Branson and Mehl. Oral view, hypotype, U.S.N.M. 114831.

FIGURE 12.—*Polylophodonta confluens* (Ulrich and Bassler). Oral view, hypotype, U.S.N.M. 114849.

FIGURE 13.—*Palmatolepis perlobata* Ulrich and Bassler. Oral view, hypotype, U.S.N.M. 114830.

PLATE 5.—REPRESENTATIVE CONODONTS OF THE HOUY FORMATION

All magnifications ×26. All specimens except Figures 3–6 from USGS coll. 9264-C (Pl. 1, section 10). Elsewhere in the United States, Figure 1 is characteristic of zone IV; Figure 2 of zones IV and V; Figures 3–6 and 8–11 of zone VI; and Figure 7 of zones V and VI. Illustrated specimens selected for quality, although those illustrated by Figures 1 and 2 are considered to have been reworked. All but Figures 3 and 6 are from the same locality.

FIGURE 1.—*Spathognathodus disparilis* (Branson and Mehl). Oral view, hypotype, U.S.N.M. 114854.

FIGURE 2.—*Spathognathodus inornatus* (Branson and Mehl). Lateral view of inner side, hypotype, U.S.N.M. 115121.

FIGURE 3.—*Pinacognathus profunda* (Branson and Mehl). Lateral view of inner side, hypotype, U.S.N.M. 114977; near section 4 of Plate 1; from exposure now covered by fill from road cut made in 1950. USGS coll. 8653-C.

FIGURE 4.—*Elictognathus lacerata* (Branson and Mehl). Lateral view of inner side, hypotype, U.S.N.M. 127855.

FIGURE 5.—*Siphonodella duplicata* (Branson and Mehl). Oral view, hypotype, U.S.N.M. 115132.

FIGURE 6.—*Polygnathus inornata* E. R. Branson. Oral view, hypotype, U.S.N.M. 115135; near section 4 of Plate 1; from exposure now covered by fill from road cut made in 1950. USGS coll. 9008-C.

FIGURE 7.—*Spathognathodus aciedentatus* (E. R. Branson). Lateral view of inner side, hypotype, U.S.N.M. 115148.

FIGURE 8.—*Siphonodella duplicata* (Branson and Mehl) var. A. Oral view, hypotype, U.S.N.M. 115140.

FIGURE 9.—*Siphonodella lobata* (Branson and Mehl). Oral view, hypotype, U.S.N.M. 115150.

FIGURE 10.—*Pseudopolygnathus prima* (Branson and Mehl). Oral view, hypotype, U.S.N.M. 114940.

FIGURE 11.—*Siphonodella quadruplicata* (Branson and Mehl). Oral view, hypotype, U.S.N.M. 115146.

Marble Falls Limestone, also needs to be better understood before its nomenclatural disposal can be satisfactorily resolved.

The deposits which remain for consideration in this paper, then, are those—clustering around the Devonian-Mississippian boundary—that constitute the Houy Formation.

The authors wish to acknowledge frankly that they are still very much puzzled about the proper relationships of these beds and faunas even though they have seen a probably very high percentage of the available data on them. Nevertheless, the stratigraphic nomenclature is one of convenience; the interpretations are subject to revision when better evidence is found.

General Features of the Houy Formation

The general features and age relations of the Houy Formation are shown in Plate 1; a view of the type section in Figure 1 of Plate 2; and views of various units in Plates 2 and 3. Its more complete sections characteristically include, in upward succession, (A) 1–3 feet of chert breccia, the Ives Breccia Member of mainly Late Devonian age (Pl. 2, figs. 3–5; Pl. 3, fig. 6); (B) 2–15 feet of black, fissile shale, the Doublehorn Shale Member of younger Late Devonian and possibly earliest Mississippian age (Pl. 2, fig. 2; Pl. 3, figs. 1, 3–5); and (C) 2 feet or less of phosphoritic material of Early Mississippian and, at places, Late Devonian age (Pl. 2, fig. 2; Pl, 3, fig. 1). These deposits show abrupt variations in thickness and associations. Their maximum observed surface thickness, however, is 17 feet, and their geographic distribution has a broadly consistent pattern.

Lenses of chert breccia (and locally conglomerate) referred to the Ives Breccia Member have been found along the eastern, northern, and western sides of the Llano region, but the Doublehorn Shale Member crops out only along the eastern side of the region, and the most westerly known occurrence of the upper phosphatic beds is along San Saba River at the eastern edge of McCulloch County. Overlapping Cretaceous deposits conceal the possible outcrop belt of the Houy Formation in the southwestern part of the region. It is not known whether the apparent absence of the Houy in mapped portions of the Riley Mountains 11 miles southeast of Llano is due to nondeposition or later removal, but possibly the center of the Llano uplift was insular in Late Devonian and earliest Mississippian time.

The lowest and presumably oldest deposits of the Houy Formation are beds or pockets of siliceous limestone with chert fragments, or silty calcareous shale that occur locally beneath the Ives Breccia Member (Pl. 1, section 11).

Next in order, and in most places the basal unit of the formation, is the Ives Breccia Member, rarely as much as 3 feet thick (Pl. 2, fig. 3). The chert fragments or nodules that constitute the bulk of the breccia at most places are locally derived, are freshly fractured or preserve the form of nodules, and are evidently little-moved lag breccias that accumulated near or at the source of supply in low places on the floor of the invading sea. In some places the breccia fragments thin out downward to be replaced by sandstone (Pl. 1, section 2) or siliceous limestone (Pl. 1, section 1). Where the Doublehorn Shale Member is missing but both Ives-type breccias and phosphatic beds are present these are likely to be inseparably mixed (Pl. 1, sections 4, 10, 12).

The Ives Breccia Member is the only unit of the Houy Formation that has previously been classified in a formal arrangement of names. Plummer considered it to be a member of the Mississippian Chappel Formation, both in his original description (*in* Bullard and Plummer, 1939, p. 15) and in his final publication (Plummer, 1950, p. 27).[1] Cloud and Barnes (1948, p. 44, 48) also considered the Ives to be Mississippian, but distinct from the Chappel Limestone and of formational rank. Although they recognized its intimate association with probable

[1] Plummer's Espey Creek Limestone Member of the Chappel Formation (Plummer, 1950, p. 22–23, 26, 28) is the same as the Chappel Limestone (Sellards, 1933, p. 92; Cloud and Barnes, 1948, p. 49–52), excluding the basal 15 inches referred by him to the Chappel at Espey Creek (Pl. 1, section 6). Plummer's Whites Crossing Coquina Member (Plummer, 1950, p. 24–26, 28), also referred by him to the Chappel Formation, is an echinodermal limestone facies in the lower part of the Barnett Formation (Cloud and Barnes, 1948, p. 45, 56–57). The "King Creek marl member" of Plummer's Chappel Formation (Plummer, 1950, p. 23, 26, 27) is an argillaceous limestone that has been surely identified only at its type site on King Creek (Pl. 1, section 3),

Devonian as well as with Mississippian rocks, they excluded from it similar deposits of known early Late Devonian age and overstressed its Mississippian aspects. New stratigraphic evidence partly summarized in Plate 1 shows Ives-type breccias beneath deposits of demonstrable Devonian age at a number of localities and confirms the earlier conclusion by Hass from conodont studies that chert breccias of this type occur at two or more levels within the Upper Devonian and possibly also in highest Middle Devonian. In fact, all the now unquestioned Ives outcrops that are also unquestionably in normal position and unequivocally dated have turned out to be Devonian. At several other places, however, the field relationships indicate either an early Mississippian age or a mixing with or infiltration by Mississippian deposits after the original deposition of the breccia (Pl. 1, sections 3, 4, 10, 12). This new evidence also leads to the suppression of the Zesch Formation of Barnes *et al.* (1947, p. 137–139), for it is a partial synonym of the Ives Breccia.

The principal unit of the Houy Formation, the Doublehorn Shale Member (Pl. 1, type section 8), is a black, fissile, radioactive (about 0.01 per cent equivalent uranium), spore-bearing shale from which, at places, large silicified pieces of the wood of *Callixylon* (identified by R. A. Scott, U. S. Geological Survey, letter of April 6, 1956, to Cloud) have weathered free. As *Callixylon* is generally considered distinctive of the Upper Devonian, its discovery reinforces the evidence of the conodonts that the black shale is in large part of Late Devonian age. At some places, however, it also includes Mississippian rocks (Pl. 1, section 7). Its maximum thickness is 15 feet, and it thins to disappearance.

The upper or phosphoritic unit of the Houy Formation, 2 feet or less thick, contains fish bones that have been determined by Dunkle and Wilson (1952) to be *Dinichthys* cf. *D. terrelli* Newberry and an unidentifiable arthrodire. *D. terrelli* and its relatives seem to be restricted to the Ohio Shale of Late Devonian age and its equivalents. Devonian conodonts have been found also in some of the phosphatic beds (Pl. 1, sections 5 and 11). The same or closely associated beds, however, contain *Sedenticellula* aff. *S. hamburgensis* (Weller) and other brachiopods considered indicative of earliest Mississippian (Cloud and Barnes, 1948, Pl. 44, figs. 12–16; *cf.* Plummer, 1950, Pl. 5, figs. 13 a–b) as well as Early Mississippian conodonts. Thus the phosphoritic beds also seem to be partly Devonian and partly Mississippian, or to include extraneous fossils from one or the other source. Some beds are radioactive. The equivalent uranium content of random samples from section 6 (Pl. 1) was determined as 0.01–0.006 per cent (Personal communication, B. A. McCall, U. S. Geol. Survey, May 17, 1956). Another sample, from a thicker zone at section 5 (Pl. 1), contained 0.02 per cent equivalent uranium (Personal communication, Roosevelt Moore, U. S. Geol. Survey, July 18, 1956).

Although phosphoritic or highly phosphatic beds seem everywhere to be lower Kinderhook, uppermost Devonian, or both, scattered phosphatic pellets are likely to occur anywhere in the Houy Formation, including the Ives Breccia Member. No appreciable amount of phosphate has been seen by the authors in the overlying Chappel Limestone (Pl. 3, figs. 2, 3–5) or elsewhere in central Texas below the Barnett Formation (Pl. 3, fig. 1) except in the Houy Formation and a disappearingly thin zone at the base of the Stribling Formation.

It is, nevertheless, evident that the deposits included in the Houy Formation comprise several distinctive lithic types and may well be an unnatural agglomeration of discrete microunits. It is generally possible, also, through laboratory studies of the conodonts found in them, to assign a probable age to any given thin interval, bed, or pocket of the Houy Formation. The separate occurrences which have not, how-

where it constitutes a few inches of the Houy Formation. Although it contains conodonts and megafossils of early Kinderhook sorts, the type "King Creek marl" occurs beneath the Ives Breccia Member (Pl. 3, fig. 6), which has here yielded only Devonian types of conodonts. Indeed, the numerous crack fillings, collapse structures (Pl. 3, fig. 2), and otherwise displaced deposits of the region are probably responsible for the mixture of "Chappel" fossils illustrated by Plummer (1950, Pl. 5). These include forms characteristic of the Lower Devonian (Plummer, 1950, Pl. 5, figs. 10, 11, 16, 21, 22), the lower or middle Kinderhook (Figs. 13a–b), and post-Kinderhook Mississippian (Figs. 1, 7–9). The upper Kinderhook (and partly ?lower Osage) Chappel Limestone proper is represented most distinctively by Plummer's Figures 23a–b of *Shumardella obsolens* (Hall).

ever, been classified to the consistent satisfaction of the authors in the field are too closely related and too small to map separately at a useful quadrangle scale. Practical considerations thus leave no satisfactory alternative to a "lumping" formational name for the entire transition complex—characterized lithically by phosphate content and radioactivity that is higher than usual for the area.

Conodont Zones and Correlation

The Houy Formation is the representative in central Texas of the widespread Upper Devonian and Lower Mississippian black-shale succession that includes such well-known eastern and midcontinent deposits as the Chattanooga, Ohio, New Albany, and Woodford shales. The basal beds of this succession are low Upper Devonian (or possibly high Middle Devonian), and the topmost beds are Mississippian (lower Kinderhook). Six conodont zones have been recognized within this sequence (Hass, 1947; 1956a; 1956b). Although these zones are transitional to one another in the boundary intervals, their modal characteristics are generally well defined, and all except one (zone V of this report) have been recognized over such a large area as to minimize the probability of significant overlap between them.

Although four of these zones have been recognized in the Houy Formation (Pls. 1, 4, 5), this zonation is not so well defined in the Llano region as elsewhere, presumably because individual beds of the Houy are commonly thin, discontinuous, and mixed. As a result, many collections from the Houy contain, in association, conodonts that are segregated at two or more stratigraphic levels where the black-shale succession is thick and well developed, as in central Tennessee and central Ohio. For present purposes these conodont zones are designated I to VI in order of upward succession.

The evidence for the zone assignment of beds in the sections illustrated in Plate 1 is not presented in detail in this paper. However, numerous collections were studied, and faunal lists were prepared for the collections that came from the sections listed in Plate 1. On that plate, these collections and the appropriate zone symbol serve as control points within the Houy Formation. Nearly all these collections contain some of the pertinent index conodonts illustrated on Plates 4 and 5 or listed in the discussion that follows.

ZONE I: The oldest beds of the Houy Formation are considered to be early Late Devonian (or possibly late Middle Devonian). They include the lower half of the Ives Breccia Member (formerly the Zesch Formation) at section 1, the entire Ives at sections 8, 9, and 11, and the pocket filling beneath the Ives at section 11. Conodonts characteristic of this zone include *Polygnathus linguiformis* Hinde, *P. pennata* Hinde, and *Neoprioniodus alatus* Hinde (Pl. 4, figs. 1–3). These species have been found in the Genundewa Limestone Lentil of the Geneseo Shale of western New York, the base of the Dowelltown Member of the Chattanooga Shale along the Eastern Highland Rim of central Tennessee (Hass, 1956b), and the lower part of the Dowelltown Member of the Chattanooga Shale in north-central Tennessee. One or more of them are also in the basal portion of the middle division of the Arkansas Novaculite in Hot Spring County, Arkansas (Hass, 1956a) and the basal beds of the Woodford and New Albany shales (Hass, 1956b).

ZONE II: Collections from loose slabs indicate that beds slightly younger than those assigned to zone I are in the Doublehorn Shale Member of the Houy Formation at sections 7, 8, and 9. Zone II conodonts are also present in the basal phosphoritic beds of the Houy at section 6. The conodonts representative of this zone include *Ancyrognathus euglypheus* Stauffer, *Palmatolepis subrecta* Miller and Youngquist, *P. marginata* Stauffer, and *P. unicornis* Miller and Youngquist (Pl. 4, figs. 4, 5). Although *P. unicornis* in some areas ranges into zone I, it is more commonly found in association with zone II species. All the species mentioned have been found in the Dowelltown Member of the Chattanooga Shale of central Tennessee (Hass, 1956b). Some of them have also been recognized in the Olentangy Shale of Ohio, a restricted part of the middle division of the Arkansas Novaculite at Caddo Gap, Arkansas, and Woodford Shale in Oklahoma.

ZONE III: Zone III is represented by the upper half of the Ives Breccia Member at section 1,

and by all the Ives at sections 2 and 3. This zone is also in the Doublehorn Shale Member at sections 5, 7, and 11. Conodonts characteristic of zone III include *Ancyrognathus bifurcata* (Ulrich and Bassler), *Palmatodella delicatula* Bassler, seven species of *Palmatolepis* (of which six are illustrated on Pl. 4), *Polylophodonta confluens* (Ulrich and Bassler), and *Neoprioniodus mutabilis* Branson and Mehl (Pl. 4, figs. 6–13). These conodonts occur in the lower faunal zone of the Gassaway Member of the Chattanooga Shale of central Tennessee (Hass, 1956b). Some or all of them have been recognized in the Gowanda Shale Member of the Perrysburg Formation and overlying Upper Devonian beds of western New York; in restricted parts of the Ohio, New Albany, and Woodford shales; and in the middle division of the Arkansas Novaculite.

ZONES IV AND V: Zone IV, which has been recognized in the Cleveland Shale Member of the Ohio Shale and the upper faunal zone of the Gassaway Member of the Chattanooga Shale, and zone V, which is tentatively based on the restricted occurrence of a few species in the basal Bedford Shale of Ohio, have not been recognized in the Houy Formation. However, some Houy collections have yielded specimens, possibly reworked, of *Spathognathodus disparilis* (Branson and Mehl), considered characteristic of zone IV; *S. inornatus* (Branson and Mehl), considered characteristic of zones IV and V; and *S. aciedentatus* (E. R. Branson), considered characteristic of zones V and VI (Pl. 5, figs. 1, 2, 7). If the time ranges of these species were the same in central Texas as in other areas, their failure to segregate zonally suggests either subsequent erosion or total by-passing of the sediments with which they would be normally associated. The alternative interpretation—that zone III conodonts in central Texas persisted through the time interval of zones IV and V—would be a more attractive one if the local sequence were thicker and less interrupted.

ZONE VI: The youngest beds of the Houy Formation are of Early Mississippian (early Kinderhook) age. They are present at sections 3, 4, 6–8, and 10–12. The phosphorite at section 5 may also belong with this group of deposits. All these collections contain an apparently mixed association in which early Kinderhook and Late Devonian conodonts of zone III predominate: In most of these collections, Mississippian specimens outnumber those from the Devonian, but in the phosphorite at section 5 Devonian species far outnumber the Mississippian representatives. Some of the conodont species which, as a group, characterize zone VI are: *Dinodus fragosus* (E. R. Branson), *Elictognathus bialata* (Branson and Mehl), *E. lacerata* (Branson and Mehl), *Pinacognathus profunda* (Branson and Mehl), *Polygnathus communis* Branson and Mehl, *P. inornata* E. R. Branson, *Pseudopolygnathus prima* Branson and Mehl, and five species of *Siphonodella* (Pl. 5, figs. 3–6, 8–11). Some or all of these have been recognized in the Sunbury Shale of Ohio and eastern Kentucky, Campbell's Falling Run Member of his (1946) Sanderson Formation and his Henryville Formation of Indiana, the Bushberg Sandstone Member of the Sulphur Springs Formation and the Hannibal Shale of Missouri, the topmost beds of the Chattanooga Shale of northeastern Oklahoma, the topmost beds of the Woodford Shale of Oklahoma, in beds near the top of the middle division of the Arkansas Novaculite at Caddo Gap, Arkansas, and the greater part of the Maury Formation of Tennessee (Hass, 1956b).

REFERENCE SECTIONS

General Statement

The best-exposed and most nearly complete section of the Houy Formation is at the juncture of Burnam Branch with Doublehorn Creek in an eastward extension of the R. M. Burnam ranch (Fig. 1B; Pl. 1, section 8; Pl. 2, fig. 1). A more conveniently accessible and by now well-known section (Barnes, 1953; 1956, p. 21–23) is that on Doublehorn Creek, just downstream from Houy Branch, on the Rubin Houy ranch (Fig. 1A; Pl. 1, section 9). The best section, at Burnam Branch and Doublehorn Creek, is chosen as the type, but the formational name is taken from the reference section on the Houy ranch. Both are in southeastern Burnet County, Texas. Faunal data and stratigraphic relationships of both sections and similar information for a selection of compar-

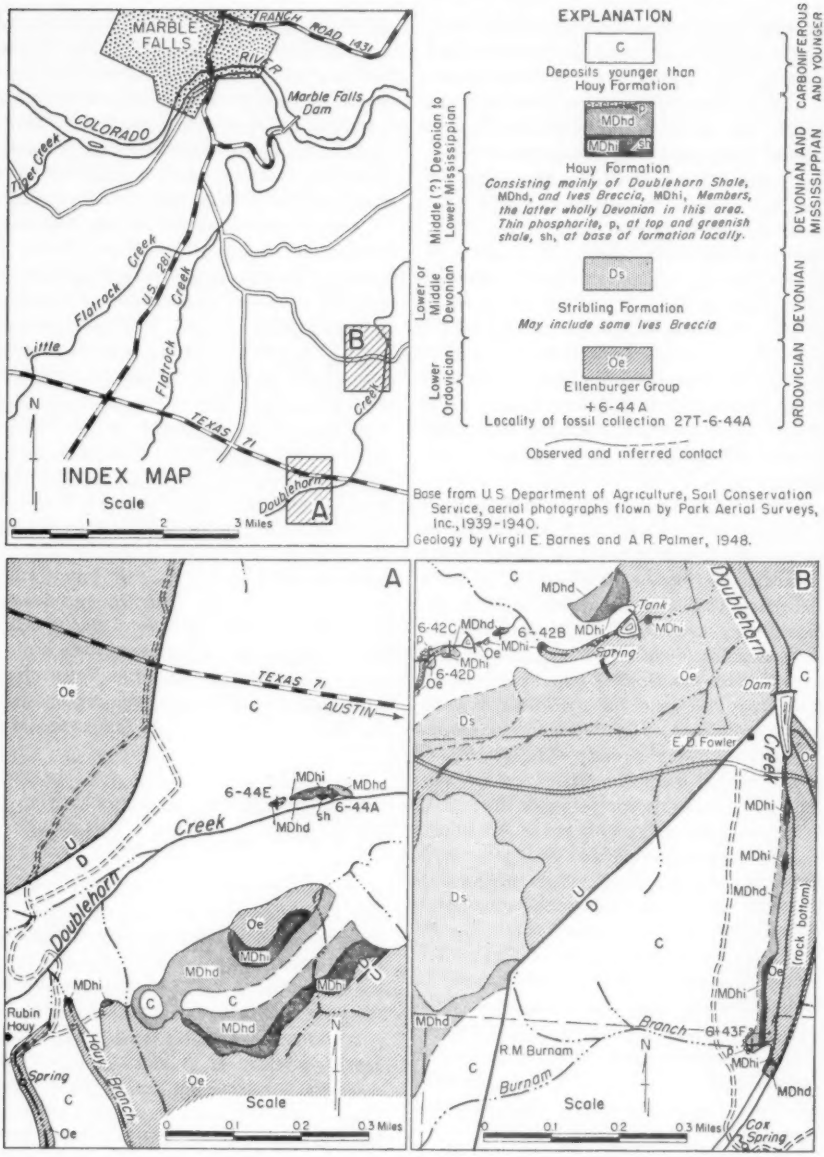

FIGURE 1.—GEOLOGIC MAPS ALONG DOUBLEHORN CREEK, SOUTHERN BURNET COUNTY, TEXAS

able sections around the Llano region are summarized in Plate 1.

Burnam Branch

The type section of the Houy Formation and the Doublehorn Shale Member is along Burnam Branch where it enters Doublehorn Creek. This is field locality 27T-6-43F[2] on Figure 1B. The following description, and section 8 on Plate 1, is supplemented by data from a submerged sequence in the bed of Doublehorn Creek.

Mississippian
- Barnett Formation (20 feet ±)—soft, somewhat fissile, punky-weathering shale with intercalated dark-brownish-gray limestone beds and lenticular concretions 1–2.5 inches thick. Contains *Lingula*, *Orbiculoidea*, *Leiorhynchus carboniferum* Girty, and conodonts of middle Mississippian (Meramec) age (Hass, 1953). Terminates upward against overhanging 4-foot ledge of black, spiculiferous Marble Falls Limestone of Early Pennsylvanian age.
- Chappel Limestone (0.3 feet)—gray, inequigranular limestone with scattered pelmatozoan columnals and upper Kinderhook conodonts. Crops out below water in natural pool on Doublehorn Creek, but missing in section up Burnam Branch.

Devonian and Mississippian
- Houy Formation (17.5 feet ±)
  - Phosphoritic member
    - Brown phosphorite containing bone fragments. USGS coll. 15552-C has abundant Kinderhook (zone VI) and a few Upper Devonian conodonts..... 0.2 feet
    - Phosphorite similar to above, in laminae 0.1–0.5 inches thick..... 0.2 feet
  - Doublehorn Shale Member—mainly black, fissile, dolomitic, spore-bearing shale, of which random samples were determined by James Schopf to have equivalent uranium content of 0.011 (coll. 3927-SD)–0.008 per cent (USGS coll. 3926-SD). A semiquantitative trace-elements analysis of a sample from USGS coll. 3927-SD was published by H. A. Tourtelot (1956, p. 76) under the number 116639. Upper foot includes some punky-weathering, silty-appearing, lighter-colored, less-fissile shale than seen below. USGS coll. 3926-SD contains Upper Devonian conodonts of zone III, among which *Palmatolepis glabra* is especially abundant. Conodonts of zone II were also found in a float slab (USGS coll. 3927-SD), and a loose piece of silicified wood was identified by R. A. Scott as the Upper Devonian genus *Callixylon*... 15 feet
  - Ives Breccia Member—coarse angular chert breccia and fractured chert, with siliceous and calcareous matrix and containing round, pea-sized phosphatic pellets. Contains low Upper Devonian conodonts (zone I) in USGS colls. 3928-SD and 3929-SD.. 2 feet

Houy Ranch

The principal supplementary section of the Houy Formation in on the Rubin Houy ranch, along Doublehorn Creek just downstream from Houy Branch and about 1.5 miles south-southwest from the type section. To reach this site from the intersection of U. S. Highway 281 and Texas Highway 71 drive 2.9 miles east, turn south through a gate 0.3 mile to the northwest bank of Doublehorn Creek and walk downstream (east-northeast) about 300 yards.

At this place (Fig. 1A, loc. 27T-6-44A) the following section is partially exposed:

Mississippian
- Barnett Formation (16 feet ±, lower 7 feet well exposed)—dark-brown to gray petroliferous shale and gray calcareous shale with thin chert beds, small turbinate rugose corals in calcareous shale about a foot above the base, compressed *Leiorhynchus carboniferum* Girty, and *Orbiculoidea* sp. on bedding surfaces of chert, and middle Mississippian (Meramec) conodonts in the shale (Hass, 1953).
- Chappel Limestone (2 feet)—medium to dark-gray inequigranular limestone with scattered pelmatozoan columnals, small rare brachiopods and trilobites, and conodonts of Kinderhook (Chouteau) age.

Devonian
- Houy Formation (10–13 feet ±)
  - Doublehorn Shale Member (type section of new unit)—black, fissile, spore-bearing shale (mainly covered) of which a random sample was determined by James Schopf of the USGS to have 0.008 per cent equivalent uranium. USGS coll. 3930-SD, from float, contains Late Devonian (zone II) conodonts 5–8 feet ±
  - Ives Breccia Member
    - Massive lag-breccia consisting mostly of locally derived large (several inches) fragments and unbroken nodules of chalcedonic and microgranular chert such as is typical of the Stribling Formation. Large angular blocks of Lower Ordovician dolomite of the Honeycut Formation are locally included within or surrounded by the chert breccia. USGS colls. 3931-SD and 3932-SD, from sparse matrix of phosphatic-siliceous debris, contain early Late Devonian (zone I) conodonts..... 3 feet

[2] This is the form of locality numbers assigned by the Bureau of Economic Geology, The University of Texas. Numbers preceded by TF are Cloud's field numbers. Others are U. S. Geological Survey permanent locality numbers. Detailed locality data are on file at the apppropriate institution. Approximate location is shown on the inset index map on Plate 1.

Yellowish-green to greenish-gray shale with inarticulate brachiopods and rare *Polygnathus linguiformis* Hinde (USGS coll. 3933-SD). Exposed by digging into stream bed beneath breccia..... 2 feet

Chert similar to that above the shale, except that it is altered to a depth of nearly an inch and in part is coated by travertine. This unit normally is beneath water level and has been seen out of water only by Barnes on September 21, 1956, when Doublehorn Creek was completely dry. Bottom not exposed..... ?

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SUBMERGED RIVER SYSTEM OF CHESAPEAKE BAY

By John T. Hack

Abstract

Logs of driving tests and borings in the deposits under the tidal estuaries in the Chesapeake Bay region have been obtained at 14 bridge sites over the Potomac, Susquehanna, Rappahannock, York, and other rivers and over Chesapeake Bay itself. The estuaries of this region are submerged river valleys cut into the Cretaceous and Tertiary sediments of the Coastal Plain, but they now contain a fill consisting of sand and gravel overlain by sandy silt of alluvial and estuarine origin. The youngest deposits are semiliquid silts (river mud) probably deposited by marine currents. Studies of ground-water supplies in terraces less than 15 feet above sea level that border some of the rivers indicate that there may be two fills of different ages separated by an unconformity and that each represents a different period of low sea level during Pleistocene time. The gradients of some river profiles steepen near the master stream, an inheritance from the period of low sea level in Pleistocene time when base level was lowered. Swampy and braided stream reaches result from the adjustment of the channel cross section to the inherited steep slope.

CONTENTS

TEXT

	Page
Introduction	817
Interpretation of borings	818
Summary of the data	820
Notes on individual crossings	822
Evidence from water wells	825
Age of valley fill	826
Profiles of the submerged Pleistocene rivers	826
References cited	830

ILLUSTRATIONS

Figure	Page
1. Index map of Chesapeake Bay	819
2. Geologic cross section at the site of the Chesapeake Bay Bridge	821
3. Geologic cross sections of rivers in the northern part of the Bay region	822
4. Geologic cross sections of rivers in the southern part of the Bay region	824
5. Hypothetical cross sections of the Potomac River near its mouth	826
6. Longitudinal profiles of the Pleistocene Susquehanna River and some of its tributaries	827
7. Map of the lower course of Zekiah Swamp Run	829

Introduction

Chesapeake Bay, a brackish-water estuary, is a classic example of a drowned river system. Several of the largest rivers of the Eastern Seaboard, including the James, Potomac, and Susquehanna, drain into it. The bay is about 25 miles wide at its widest point and 175 miles long from the Atlantic Ocean at Cape Henry to Havre de Grace at the mouth of the Susquehanna River. The bay is shallow and generally less than 60 feet deep even in the deepest places. Much of the bay floor is underlain by mud deposited by either tidal or turbidity currents. Beneath the mud are Pleistocene and Recent sediments that in some places are more than 100 feet thick. The talweg, or line connecting the lowest points on the surface of the bedrock, outlines an old valley system, which near Annapolis is 200 feet below sea level. The valleys are in pre-Pleistocene deposits consisting for the most part of unconsolidated sand, silt, clay, and shell marl. The Potomac Estuary and the upper Chesapeake Bay crosscut Eocene and upper Cretaceous sediments. The estuaries in eastern Maryland and in Virginia are cut primarily in Miocene sand, clay, and diatomaceous earth.

The geology of the Quaternary sediments in the bay and in the estuaries tributary to it can be inferred from the logs of borings that have penetrated these sediments. Such logs of borings made to test foundation conditions at bridge sites or proposed bridge sites have been examined by the writer for 14 localities. The data are sufficiently clear and consistent from one locality to another that the history of deposition can be inferred with some confidence.

Subsurface information used in this report has been obtained from the Maryland State Roads Commission, J. E. Greiner and Co. of Baltimore, the Corps of Engineers, U. S. Army, and the Department of Highways, Commonwealth of Virginia. The writer wishes to thank Mr. Carl W. A. Supp, J. E. Greiner and Co., and Mr. W. A. Parrott, Virginia Department of Highways, for their suggestions and help in obtaining data, and H. F. Ferguson, G. M. Richmond, and J. F. Smith, members of the staff of the U. S. Geological Survey, for critical reading of the manuscript.

Interpretation of Borings

The interpretation of the geology of the Quaternary fill is based primarily on an examination of logs of borings. Samples of only one boring, at the Rappahannock River (Fig. 1, loc. 10), were examined by the writer. A careful geologic interpretation based on microscopic examination of samples, comparison with well logs on the adjacent land, and some paleontologic evidence has been supplied by J. E. Greiner and Co. (Mr. Carl W. A. Supp, oral communication) at the site of the Chesapeake Bay Bridge. A geologic cross section at this site has been published by the Maryland Department of Mines and Water Resources (Ryan, 1953). The data on boreholes at these two localities have been used as a control for the interpretation of that from other localities.

The method of boring used at the bridge sites consists of driving a casing or pipe with a power hammer. The borings are from a few feet to more than 400 feet deep and at the large river crossings penetrate the Pleistocene deposits into the tertiary or Cretaceous sediments. Samples are taken at intervals of about 2–15 feet. Color and texture are the principal characteristics logged and for the purposes of this paper are significant. The greenish color of the glauconitic sediments of the Cretaceous and Eocene, for example, may be sufficient to distinguish them from the overlying brown or gray Pleistocene deposits. Principal reliance, however, is placed on texture. In the logs for the Chesapeake Bay Bridge, information on texture is detailed and includes percentage of sand, silt, and clay in the samples. At other sites the textural information is merely a brief word description such as "sand and gravel with traces of clay," or "sandy mud." The consistency of the material is generally recorded, whether soft, hard, or crusty, or the casing may be refused. Unusual materials such as shells, wood fragments, peat, and concretions are noted.

Particularly useful is the record of the number of hammer blows required to drive the casing, sampler, or test pile a given distance. Contacts between deposits of markedly different textures, such as that between the gravel at the base of the Pleistocene and the finer fill above it, in places are clearly recorded on the logs by a sharp increase in the number of hammer blows.

But the most helpful criterion by which the contact between Pleistocene and older beds and consequently the outline of the Pleistocene valley walls may be recognized is simply the presence or absence of gravel. Tertiary deposits of this part of the Coastal Plain in which the valleys are cut contain almost no gravel or coarse sand. This is well known from surface exposures as well as from data obtained from water wells. Some layers of fine gravel occur in the Yorktown Formation of Miocene age, but they are rare. The Potomac Group of Cretaceous age contains gravel, but its outcrop area is localized along the western margin of the Costal Plain. Therefore irregular lenses of gravelly and coarse sand deposits encountered by the borings at bridge sites generally indicate valley fill. Furthermore the gravel in the fill is commonly localized at the base of the fill along the unconformity between the fill and the more regularly bedded Tertiary deposits.

Samples from one of the deep borings at the

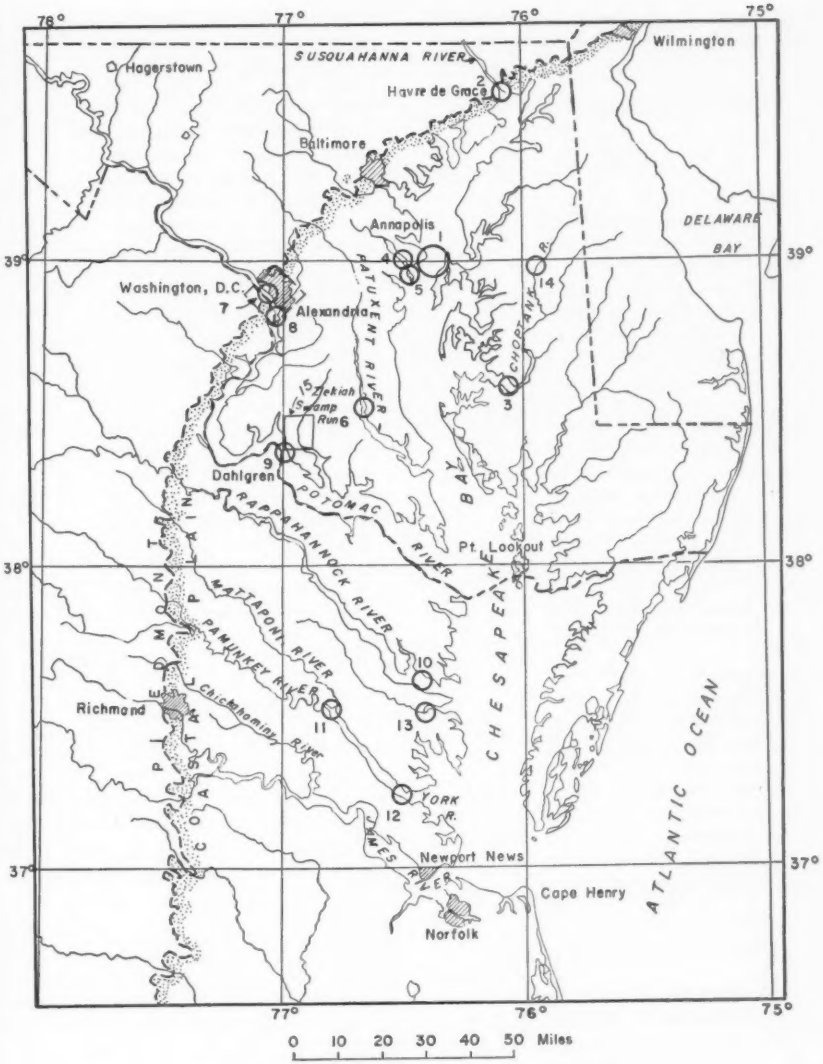

FIGURE 1.—INDEX MAP OF CHESAPEAKE BAY

Shows localities at which borehole records have been obtained. Numbers refer to text and profiles in Figures 2, 3, and 4.

site of the proposed crossings over the Rappahannock River (Figs. 1 and 4, loc. 10) were made available by the Virginia Department of Highways, and selected samples were examined by the writer. Foraminifera contained in the samples were identified by Ruth Todd, U. S. Geological Survey. The samples were examined by I. G. Sohn of the U. S. Geological Survey for Ostracoda but were barren. A stratigraphic section of the material in this borehole follows:

Geologic section at hole number 4 Rappahannock River crossing, Grey's Point to Whitestone (from logs of borings and megascopic examination of dried samples)

Altitude of top of section: mean sea level

	Thickness (feet)	Depth (Cumulative to base of unit)
1. Water	30.5	
2. Unit D: Semiliquid muck, casing settles of its own weight. No sample obtained	20.0	50.5
3. Unit C: Silt, light olive gray (5Y 6/1, dry);[1] contains clear crystals of selenite, which presumably formed while the samples were drying, and small flakes of muscovite; scattered, dark yellow-orange stains (10YR 6/6). Flakes of organic matter in lower part. Sample at top of unit contains Recent Foraminifera indicative of shallow estuarine conditions (Ruth Todd, personal communication). These are *Ammobaculites* cf. *A. exilis* Cushman and Bronnimann, *Elphidium incertum* (Williamson), and *Globigerina* sp.	24.0	74.5
4. Silt, light olive gray (5Y 6/1). Layers of peat and flakes of organic matter. Minute flakes of muscovite common; some very fine quartz sand and grains of glauconite. Few selenite crystals that may have formed as samples were drying	20.0	94.5
5. Unit B: Sand, silty and sandy silt, light olive gray to olive gray (5Y 6/1–5Y 4/1). Scattered flakes of muscovite and grains of glauconite. Traces of organic matter	20.0	114.5
6. Unit A: Sand, yellowish gray (5Y 7/2), fine-grained, chiefly .125–.25 mm in diameter. Most angular to subangular, but some rounded and polished grains. Few grains of glauconite	13.0	127.5
7. Sand, very pale orange (10YR 8/2), coarse: and medium-grained. This unit is interpreted as the base of the Pleistocene beds	12.0	139.5
8. Miocene sediments: Silt, light olive gray (5Y 6/1) and layers of very fine-grained sand. Pale yellowish-orange (10YR 6/6) and dark yellowish-orange (10YR 6/6) stains. Some coarse sand grains, crystals of selenite that may have formed as samples were drying, and flakes of muscovite. Barren of microfossils... (Ruth Todd, I. G. Sohn, K. E. Lohman, personal communication).	34.0	173.5
9. Sand, light olive gray (5Y 6/1), very fine grained, silty and clayey. Abundant shell fragments. Has mealy feel, generally characteristic of Miocene sands, when dry. Contains Foraminifera characteristic of Miocene marine sediments (Ruth Todd, personal communication)	33.0	206.5
10. Silt, sandy, yellowish gray (5Y 8/1), and clayey very fine grained sand. Contains Foraminifera characteristic of Miocene marine sediments	24.0	230.5
Total thickness (excluding water at top)	200.0	

[1] Color symbols from National Research Council Rock Color Chart (1948).

Summary of the Data

Comparison of the lithologic sequences at the 13 river crossings with those at the site of the Chesapeake Bay Bridge indicates that the sequence of deposition and erosion has probably been similar at all localities. In each place the country rock, a Tertiary or Cretaceous sediment or—at the sites near the fall line—a crystalline rock, is commonly overlain unconformably by a thin bed of sand or gravel. This material covers the bottoms and walls of submerged valleys. It is overlain by finer-grained silt and silty sand that in places contains shells, in places plant material, either wood or peat, and locally contains lenses of gravel, though large lenses of gravel have not been formed above the base of the fill. The silty deposits are overlain by semiliquid clayey silt generally referred to in the logs as mud.

The structure of the fill is exemplified at the Chesapeake Bay Bridge site, shown in Figure 2.

UNITS A AND B: These units are for the most

part sand and gravel. The writer interpreted them as coarse-textured river-bed deposits laid down in the channels of the Pleistocene Susquehanna River and its tributaries. Driving tests indicate that this material is firm. According to J. E. Greiner and Co. the material at the Bay Bridge has a low moisture content ranging from 10 to 25 per cent (unpublished drawings).

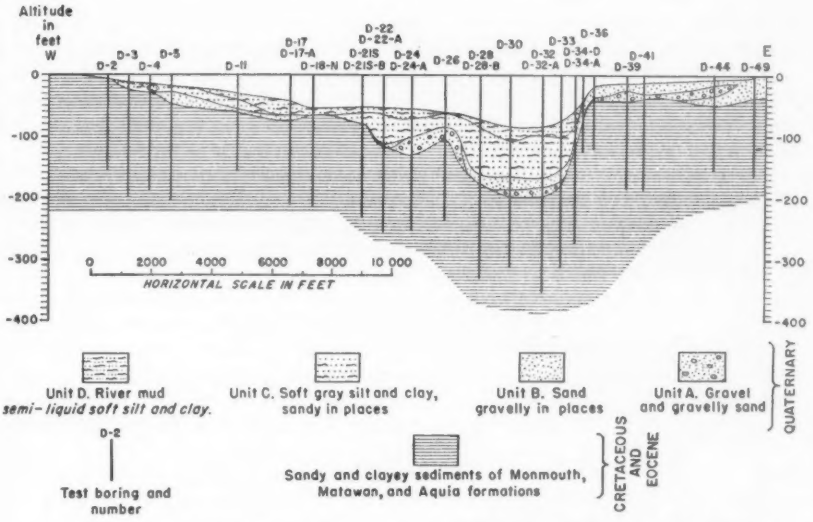

FIGURE 2.—GEOLOGIC CROSS SECTION AT THE SITE OF THE CHESAPEAKE BAY BRIDGE Between Sandy Point and Kent Island (Fig. 1, loc. 1). Modified from Ryan (1953, Pl. 2).

UNIT C. This material is silt and sandy silt laid down in most places on the fluviatile sand and gravel. It makes up the largest part of the submerged valley fill at many sites. Physically the material is considerably less resistant to the test piles than is the sandy or gravelly material beneath it. At the Rappahannock crossing, for example, the hammer blows required to drive the sampler 1 foot rose sharply in number from 5 or less to 15 or 20 at the base of unit C. Moisture content, as shown by the Bay Bridge data, is high and ranges from 60 to 130 per cent dry weight. In some places the unit is 150 feet thick. In other places, as at the York River Bridge, it is apparently lacking.

The silt and sandy silt of unit C are heterogeneous. In some places the material is sandy, in others silty or even clayey, and locally it contains lenses of gravel. The unit also contains oyster and other shells in some places and considerable organic matter such as peat and wood fragments.

The writer believes that unit C was deposited during a period of rising sea level and represents a stage intermediate between a fluviatile stage and an estuarine stage. The presence of oyster shells suggests deposition under water, whereas the peat and wood fragments suggest nearshore, tidal-marsh, or deltaic conditions.

UNIT D: This material, generally referred to as mud or river mud, is soft organic silt, which probably was deposited chiefly during Recent time but is still being deposited. It is a deposit transported by wave, tidal, or other underwater currents and may have been derived from the shore or from fresh-water streams that debouch into the estuaries. Test-pile and core data indicate that it is very infirm and has a high moisture content, in places exceeding 150 per cent dry weight. The test piles and sampling spoons penetrate it of their own weight. Foraminifera identified by Todd are of

Recent age and indicate deposition in brackish water (Ruth Todd, personal communication).

UNIT E: This material consists of a mixture of silty loam, sand, gravel, and boulders. It has presumably formed on the subaerial slope by slumping of soil material.

feet. The fill lies on the Aquia Formation of Eocene age and the Monmouth and Matawan formations of Cretaceous age.

SUSQUEHANNA RIVER AT HAVRE DE GRACE, MARYLAND (FIGS. 1 AND 3, LOC. 2): The data for this crossing at the site of the highway

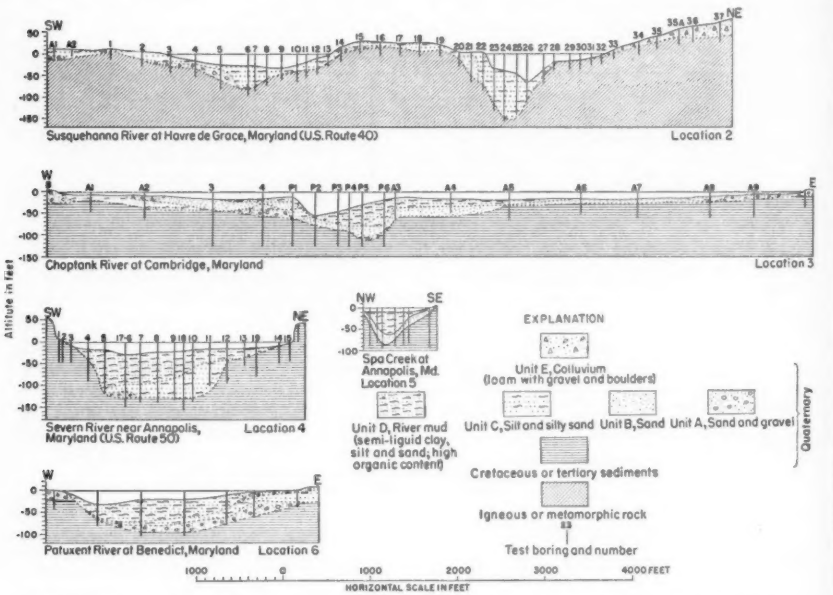
Susquehanna River at Havre de Grace, Maryland (U.S. Route 40)
Location 2
Choptank River at Cambridge, Maryland
Location 3
Severn River near Annapolis, Maryland (U.S. Route 50)
Location 4
Spa Creek at Annapolis, Md.
Location 5
Patuxent River at Benedict, Maryland
Location 6
Altitude in feet
EXPLANATION
Unit E, Colluvium (loam with gravel and boulders)
Unit D, River mud (semi-liquid clay, silt and sand; high organic content)
Unit C, Silt and silty sand
Unit B, Sand
Unit A, Sand and gravel
Quaternary
Cretaceous or tertiary sediments
Igneous or metamorphic rock
Test boring and number
1000 0 1000 2000 3000 4000 FEET
HORIZONTAL SCALE IN FEET

FIGURE 3.—GEOLOGIC CROSS SECTIONS OF RIVERS IN NORTHERN PART OF THE BAY REGION Interpreted from borings.

TERRACES: Submerged terraces cut on pre-Pleistocene materials and overlain by sands and gravel of units A and B are at many of the localities. Some may be river terraces older than the sandy and gravelly deposits in the lowest part of the fill. Some may be wave-cut terraces formed during a stage of rising sea level when sand and gravel were strewn by the action of shore currents on wave-cut benches.

NOTES ON INDIVIDUAL CROSSINGS

CHESAPEAKE BAY BRIDGE (FIGS. 1 AND 2, LOC. 1): The geology of the site of the Chesapeake Bay Bridge has been published elsewhere (Ryan, 1953, Pl. 2), and details are not repeated here. The base of the Quaternary fill at its lowest place is at an altitude of −200

bridge on U. S. Route 40 were furnished by the Maryland State Roads Commission. The interpretation shown in Figure 3 is based on the textural descriptions of the samples, as test-pile data were not obtained. The pre-Pleistocene rock is described in the logs as granite, hard rock, or rock, and the lowest point at which it was encountered was at an altitude of −140 feet.

CHOPTANK RIVER AT CAMBRIDGE, MARYLAND (FIGS. 1 AND 3, LOC. 3): The Choptank River at locality 3 is a wide tidal estuary with low shores and is crossed by a highway bridge. The borehole data were furnished by the Maryland State Roads Commission and indicate that the talweg of the submerged channel is at about −130 feet altitude. Bedrock is probably the Choptank Formation of Miocene age.

SEVERN RIVER NEAR ANNAPOLIS, MARYLAND (FIGS. 1 AND 3, LOC. 4): This crossing is at a bridge site along the alignment of the new highway between Washington, D. C., and the Chesapeake Bay Bridge, known as the Annapolis bypass on U. S. Route 50. The submerged Quaternary channel of the Severn River is cut into Cretaceous sediments, both the greensand of the Matawan Formation and the porous sand of the Magothy Formation. The presence of the Magothy Formation is shown in the deepest holes by wood, a characteristic constituent of this formation. The Pleistocene fill apparently extends to a depth of −125 feet.

SPA CREEK AT ANNAPOLIS, MARYLAND (FIGS. 1 AND 3, LOC. 5): Spa Creek is a small tidal river crossed by a bascule bridge on a road between Annapolis and Eastport. The exploration borings furnished by the Maryland State Roads Commission penetrate greenish glauconitic sand and clay, which are probably undisturbed Cretaceous deposits.

PATUXENT RIVER AT BENEDICT, MARYLAND (FIGS. 1 AND 3, LOC. 6): Test piles were driven to obtain information on foundation conditions for the new bridge between Benedict and Hallowing Point, Maryland. A geologic cross section prepared by J. E. Greiner and Co. (unpublished report) based entirely on test-pile data rather than on borings is used as a basis for the cross section shown in Figure 3. According to this interpretation the Quaternary fill rests on the Nanjemoy Formation of Eocene age at an altitude of −97 feet at its lowest point.

POTOMAC RIVER AT WASHINGTON, D. C. (FIGS. 1 AND 4, LOC. 7): The geologic cross section at locality 7 is based on a profile made in 1925 for the Arlington Memorial Bridge between Washington, D. C., at 23 Street, NW., and Arlington, Virginia. This profile was made available by the Corps of Engineers, U. S. Army. Every boring penetrates firm bedrock, probably of Precambrian age. The surface of the bedrock at this locality is described by Darton (1950, p. 25). Its average altitude is −45 feet.

POTOMAC RIVER AT ALEXANDRIA, VIRGINIA (FIGS. 1 AND 4, LOC. 8): The cross section at locality 8 is based on a line of borings made in 1941 across the river (data furnished by the Corps of Engineers, U. S. Army). The line extends across the river from the vicinity of First Street in Alexandria to Marbury Point, Maryland. The interpretation shown in Figure 4 is based on the logs of samples as well as the record of hammer blows on the casing. The Pleistocene fill reaches its maximum depth at an altitude of −100 feet. Bedrock is probably clay of the Potomac group.

POTOMAC RIVER NEAR DAHLGREN, VIRGINIA (FIGS. 1 AND 4, LOC. 9): Borehole data were examined at the site of the toll bridge on U. S. Route 301 that crosses the Potomac River from Ludlow Ferry, Maryland, to a point near Dahlgren, Virginia. The logs were furnished by the Maryland State Roads Commission. Some of the borings that reach a depth of −196 feet penetrate stratified clay of the Marlboro Clay Member of the Nanjemoy Formation of Eocene age. This member is overlain unconformably by sand and sand and gravel in the filled channel whose maximum depth is at an altitude of −165 feet.

RAPPAHANNOCK RIVER AT GREY'S POINT, VIRGINIA (FIGS. 1 AND 4, LOC. 10): The Rappahannock River is crossed by a line of test borings at a proposed bridge site only a few miles upstream from its junction with Chesapeake Bay. The submerged channel is cut into Miocene sand and silty sand containing shell marl and scattered shells of Miocene age. Maximum depth of the filled channel is about −145 feet.

PAMUNKEY RIVER AT WEST POINT, VIRGINIA (FIGS. 1 AND 4, LOC. 11): The Pamunkey River joins the Mattaponi River at West Point, Virginia, to form the York River. The geologic cross section shown in Figure 4 is based on logs of borings as well as data on driving tests furnished by the Virginia Department of Highways. The submerged valley fill lies on Miocene sediments at a maximum depth of −80 feet.

YORK RIVER AT YORKTOWN, VIRGINIA (FIGS. 1 AND 4, LOC. 12): The York River at the narrows between Yorktown and Gloucester Point is crossed by a highway bridge on U. S. Route 17. Data on this site, furnished by the Virginia Department of Highways, indicate that the lowest part of the submerged fill is on the Yorktown Formation of Miocene age at an altitude of −125 feet.

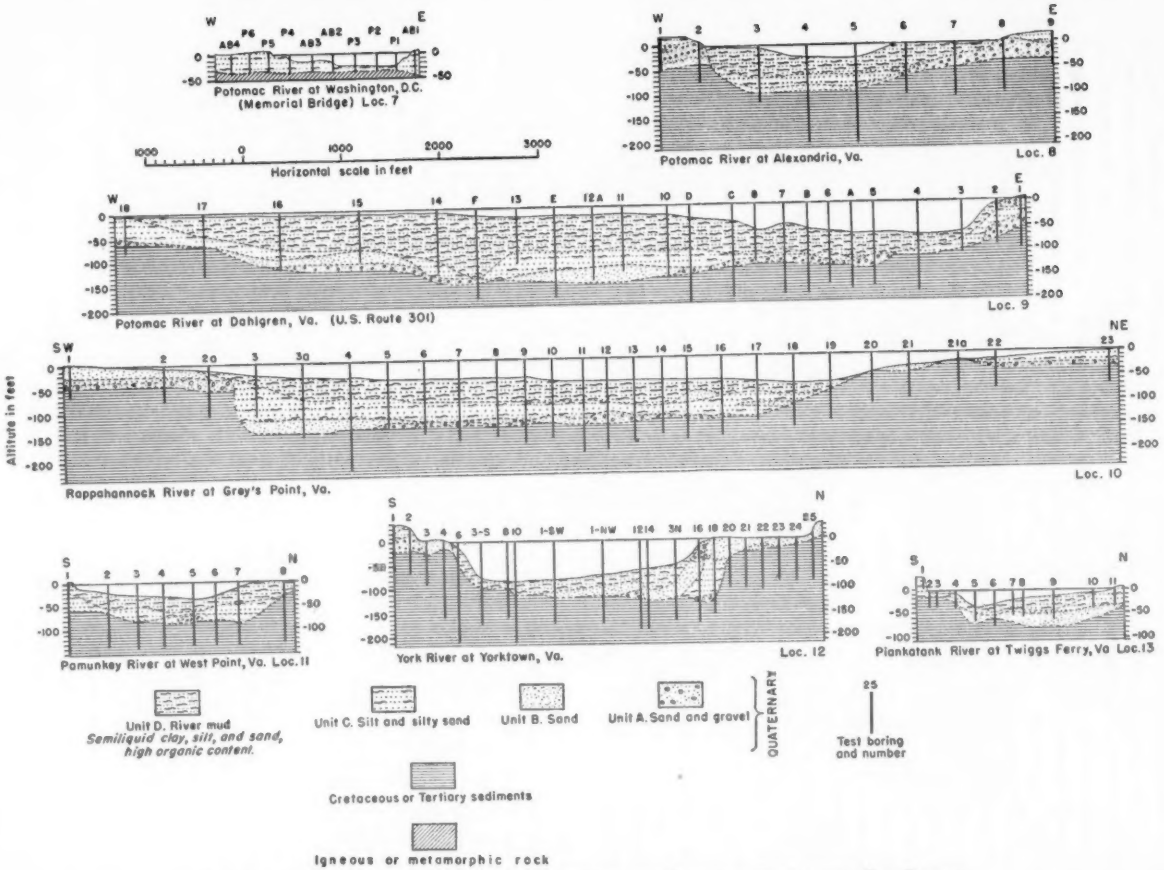
Potomac River at Washington, D.C. (Memorial Bridge) Loc. 7
Potomac River at Alexandria, Va. Loc. 8
Horizontal scale in feet
Potomac River at Dahlgren, Va. (U.S. Route 301) Loc. 9
Rappahannock River at Grey's Point, Va. Loc. 10
Altitude in feet
Pamunkey River at West Point, Va. Loc. 11
York River at Yorktown, Va. Loc. 12
Plankatank River at Twiggs Ferry, Va Loc. 13
Unit D. River mud
Semiliquid clay, silt, and sand, high organic content.
Unit C. Silt and silty sand
Unit B. Sand
Unit A. Sand and gravel
QUATERNARY
Test boring and number
Cretaceous or Tertiary sediments
Igneous or metamorphic rock

FIGURE 4.—GEOLOGIC CROSS SECTIONS OF RIVERS IN THE SOUTHERN PART OF THE BAY REGION
Interpreted from borings.

PIANKATANK RIVER AT TWIGGS FERRY, VIRGINIA (FIGS. 1 AND 4, LOC. 13): The Piankatank River is a small tidal estuary between the York and Rappahannock Rivers. It is crossed near its mouth by a line of borings made for the Virginia Department of Highways to test for a proposed bridge. The depth of the borings is inadequate to determine the complete thickness of the Quaternary fill. As shown in Figure 3, holes 5, 7, 8, 9, 10, and 11 penetrate only sandy sediments, which are probably Quaternary. The rock encountered at the bottom of these holes is predominately medium- to coarse-grained sand, a type considered to be uncommon in the Miocene. By comparison with the cross sections it may be inferred at other sites that the Quaternary sand layer at the bottom of the holes is not very thick and rests on Miocene beds. If this is true, the base of the fill is probably at about −60 or −70 feet.

CHOPTANK RIVER AT GREENSBORO, MARYLAND (FIG. 1, LOC. 14): Logs of borings which completely penetrate the Quaternary fill are available at the site of a bridge on a bypass around the city of Greensboro on the Choptank River. Because of the small size of this channel the cross section cannot be shown on the same scale as the other crossings and is not illustrated. The river is only 150 feet wide and 3 feet deep. The fill consists of light-gray silt and clay. At a maximum depth of about 25–30 feet it rests on softer, brown sandy silt, which contains fossil shells in places. The logs, though they cannot be interpreted as surely as at other sites where gravel deposits occur, indicate that the talweg probably lies at −25 to −30 feet.

FIGURE 4.—GEOLOGIC CROSS SECTIONS OF RIVERS IN THE SOUTHERN PART OF THE BAY REGION
Interpreted from borings.

EVIDENCE FROM WATER WELLS

Additional evidence on the geology of the submerged Pleistocene valleys has been obtained from studies of ground-water supplies in the Baltimore area by Bennett and Meyer (1952) and from studies in the area of Point Lookout, Maryland, at the mouth of the Potomac River by Ferguson (1953). In the Baltimore region, wells bordering the Patapsco River on an 8- to 10-foot terrace known as Sparrows Point penetrate more than 125 feet of Pleistocene deposits consisting of a basal gravel overlain by dark-gray clay and lenses of gravel. The clay contains plant fragments, ostracods, *Ostrea* sp., and a large thick-shelled pelecypod, *Rangia cuneata*, according to Bennett and Meyer (1952, p. 69). The sequence of deposits is roughly similar to that in the test-boring logs at the river crossings. The correlation is of particular interest because it shows that a part of the fill of the submerged valley underlies a terrace that is now 8–10 feet above sea level.

Similar evidence has been found at the mouth of the Potomac River in the vicinity of Point Lookout, Maryland. (Ferguson, 1953, p. 37, Pl. 5). In this area a terrace 3–15 feet above sea level is underlain by a sequence of Pleistocene deposits that consist of coarse sand and gravel overlain by estuarine clay in places more than 150 feet thick. A bed of marine fossils in an exposure between sea level and 15 feet above sea level at Cornfield Harbor on the terrace 3 miles north of Point Lookout along the Potomac shore, has been described in detail by Mansfield (1928, p. 129) and by Blake (1953). The fossils are in a marine clay on top of at least 125 feet of Pleistocene deposits. Ferguson's data indicate that a period of high sea level intervened between the formation of the basal gravel and the fixing of the present sea level; at the high level the sea was at least 15 feet higher than at present.

Another change in sea level is indicated by the topography at Point Lookout. Here, the terrace that is 3–15 feet above sea level is cut by drowned valleys that make small harbors close to shore. These valleys indicate that subsequent to the high stand the sea dropped to a level below the present level and later rose.

Two interpretations are possible: (1) The basal gravels at Point Lookout and Sparrows Point may be extensions of the gravel recorded along the talweg of the Pleistocene channels and may therefore be river gravels of the same age. If this is true, the evidence indicates a period of low sea level followed by a rise to over 15 feet above the present, then by a drop to a level somewhat below the present, and finally by a rise to present level. (2) On the other hand the data are not compelling evidence that the gravel at Point Lookout and the gravel at Sparrows Point are equivalent to the basal

gravels recorded in the river crossings. Perhaps we are dealing with two periods of low sea level somewhat near the same altitude. Perhaps the Pleistocene fill at Point Lookout is older than the cutting of the present bay and represents a fill still older than that recorded at the bridge crossings. Should this interpretation be correct

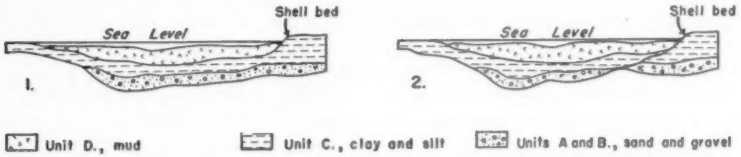

FIGURE 5.—HYPOTHETICAL CROSS SECTIONS OF THE POTOMAC RIVER NEAR ITS MOUTH

Shows two interpretations of the structure of the Pleistocene fill: (1) The fill under Point Lookout is correlative with the fill over the talweg. (2) There are two fills of different ages.

the sequence of events would be as follows: a period of low sea level (150 feet lower than the present) was followed by a rise to a level over 15 feet above the present, then by a period of low sea level when the gravel along the talweg was deposited, and finally by a rise to the present. The two possible interpretations of the evidence are illustrated graphically in Figure 5. The second hypothesis is preferred by the writer because of additional evidence to be discussed.

AGE OF VALLEY FILL

Data on the age of the valley fill are scant and indicate only that it must be Pleistocene and Recent. As the uplands that stand 200–300 feet above sea level in the Chesapeake Bay region are underlain by river gravels generally regarded as Pliocene, and lower terraces above present sea level are regarded as Pleistocene (Hack, 1955, p. 26), it may be reasonably inferred that the submerged fill is at least as young as Pleistocene.

Most geologists accept the proposition discussed by Daly (1934) that world-wide fluctuations in sea level resulted from and accompanied the fluctuations of the continental ice caps. Periods of low sea level corresponded with periods of maximum cold and greatest ice advance. Gravelly fill along the talweg of the submerged valleys of Chesapeake Bay is at altitudes far below the present sea level and was presumably deposited during an ice advance or during a stand of the continental ice sheet that corresponded to a sea level far lower than the present level.

Regardless of the possible complexities in the history of cutting and filling, the latest period of cutting recorded by the position of the talweg is probably of Wisconsin age. As shown at the bridge crossings the gravels in contact with the valley walls and floor are the only gravels of any appreciable thickness and, except possibly at locality 3, there are no gravels within the fill younger than the gravel on the talweg. The gravel on the talweg, therefore probably represents the most recent period of low Pleistocene sea level during which the now submerged valleys were occupied by fresh-water streams. If the valleys were eroded subaerially to a lesser depth after the basal gravel was deposited, younger gravel deposits within the central part of the fill at a level higher than the valley floor would be expected, for sources of abundant gravel occur in all these valleys on the terraces. No gravel deposits younger than the basal gravels are recorded.

The gravel at Point Lookout, Maryland, described by Ferguson (1953), and at Cornfield Harbor (Wailes Bluff) underlies a shell bed whose age is greater than 35,000 years from a C^{14} determination by Rubin and Suess (1956). This may suggest also that at least two periods of low sea level are involved in the cutting and filling of the Chesapeake Bay drowned valleys, for presumably at least one major ice advance and concurrent lowering of sea level took place less than 35,000 years ago.

PROFILES OF THE SUBMERGED PLEISTOCENE RIVERS

Borehole data are available at too few localities to establish the longitudinal profile along the floor of any one of the submerged

FIGURE 6.—LONGITUDINAL PROFILES OF THE PLEISTOCENE SUSQUEHANNA RIVER AND SOME OF ITS PRINCIPAL TRIBUTARIES IN THE CHESAPEAKE BAY REGION

Solid lines used above sea level where data are from topographic maps. Dashed lines used below sea level where data are available only at numbered localities shown by circles. Numbers refer to localities in Figure 1. Data on elevation of Susquehanna River bed at Conowingo Dam from Wilson, 1929.

channels. Nevertheless, since all the channels are on an interconnected system, a reasonable speculation on the nature of the profiles is permitted. The longitudinal profiles of the Pleistocene Susquehanna and some of its tributaries are reconstructed and shown in Figure 6. The positions of the junctions of the various streams have, of course, been estimated. As the Chesapeake Bay is relatively narrow, the error in such an estimate is not large, and it is immaterial whether a tributary enters the main stem at the mouth of its valley or whether it parallels the master stream on the main valley floor for some distance before joining.

The profile of the Pleistocene Susquehanna below Annapolis has been estimated, as no data are available for the control of any part of this channel below the Chesapeake Bay Bridge. The profile may be concave upward, may be straight, or may be convex upward. Unless there has been regional deformation the altitude must be lower at Cape Henry than at the Chesapeake Bay Bridge. If the Susquehanna had sufficient time during the period of low sea level to develop a normal stream profile, concave upward, one of the many possible reasonable profiles may be drawn by a method developed and described elsewhere (Hack, in press). The profile of the lower Pleistocene Susquehanna below Annapolis, shown in Figure 6, is based on the assumption that the channel follows a logarithmic curve, and that the same curve passes through the known elevations at the Bay Bridge and at Havre de Grace. This curve may be too steep.

The construction of Figure 6 shows that nearly all the tributaries have profiles that steepen as they approach the main stem of the Pleistocene Susquehanna. This steepening is very pronounced. Even if the main stem had no fall below Annapolis but maintained an elevation of −200 feet, the profiles of several of the tributaries, including the Piankatank, the Pamunkey, and the Choptank, would have to be convex upward in order to join at that elevation.

The anomalous gradients of the lower courses of the tributary streams is most readily explained as a relict feature inherited from a partial adjustment of the profiles to a low Pleistocene sea level. During the period of low sea level deep valleys were eroded in the reaches immediately upstream from the Pleistocene Susquehanna River. The downcutting worked headward various distances depending on the size of the tributary stream.

Some of the tributary streams with steepened gradients have another anomalous feature in their lower courses probably related to the Pleistocene downcutting. These streams have a braided, swampy character in their lower reaches. They are incised in steep-walled valleys, with flat floors, swampy tracts, and anastomosing channels. If the vegetation were completely removed they would probably be similar in aspect to typical braided streams.

Most of the streams shown in Figure 6 are steepened only below present sea level so that the character of the valley floor cannot be determined. Some smaller streams, however, have profiles convex upward in parts of their courses still above present sea level. One such stream that also has a braided character is Zekiah Swamp Run (Fig. 1, loc. 15), a small tributary of the Potomac, which rises on the plain of southern Maryland. Its profile is slightly steeper near sea level than it is a short distance upstream (Fig. 6). It has cut a wide, steep-walled valley more than 100 feet below the upland (Fig. 7). The land along the channel is low, swampy, and broken by numerous small channels. Tributaries enter at the valley sides and flow parallel to the main stream for a distance before joining it. It might appear at first glance that the valley of Zekiah Swamp Run is a drowned valley, swampy because it was filled by river silt during a rising sea level. Actually, however, its slope is steep, relative to the upstream part of its course, and the present analysis indicates that it is probably flowing on only a shallow fill or directly on Tertiary bedrock.

The association of a steep gradient with an abnormally wide channel of anastomosing character may be explained by the concept of adjusted cross section of stream channels discussed by Rubey (1952, p. 129). Rubey's analysis of streams indicates that where the depth-width ratio (or form ratio) is constant, the graded slope decreases with decrease of load or grain size, or with increase of discharge. The depth-width ratio like slope, is a dependent variable, and a stream may adjust to an increase in discharge, for example, by an increase in

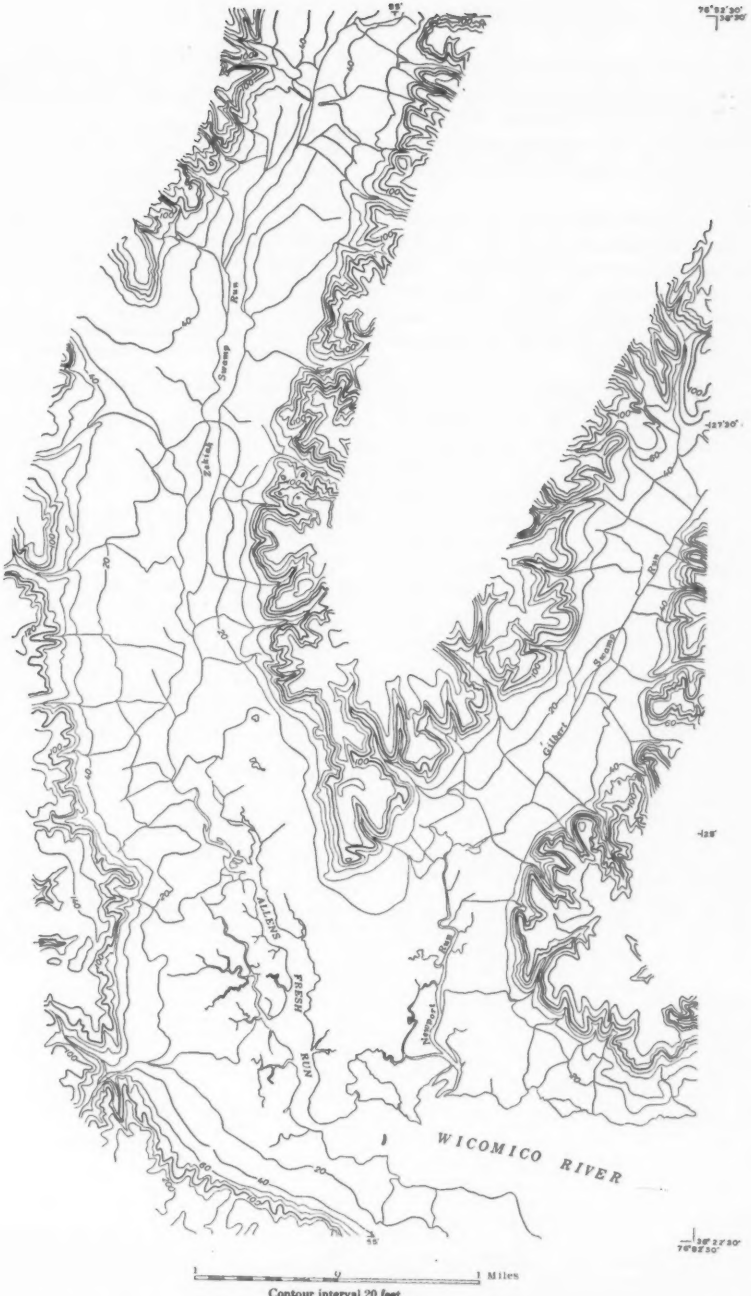

FIGURE 7.—MAP OF THE LOWER COURSE OF ZEKIAH SWAMP RUN, CHARLES COUNTY, MARYLAND
Redrawn from map of Popes Creek quadrangle (U. S. Geol. Survey, 1953).

width (in relation to depth) rather than by a decrease in slope. Rubey suggests (1952, p. 135) that the Platte River of Nebraska may have an extraordinarily wide channel cross section because it flows on a steep slope inherited from earlier streams.

Similarly it may be that some of the coastal-plain streams like Zekiah Swamp Run have a wide channel cross section as a consequence of the steepened gradient formed during the Pleistocene when base level as well as load and discharge characteristics of these streams were different.

Some larger streams, such as the Mattaponi and Chickahominy (Fig. 1), in contrast to streams like Zekiah Swamp Run have meandering patterns and narrow channels in their lower courses. These streams have gradients so low that tidal effects extend many miles upstream, and they are bordered by tidal marshes in their lower courses. Presumably such streams are flowing on Quaternary fill. The meandering pattern, as suggested by Leopold (1955), may be a consequence of a lowered gradient in this case imposed by a rising sea level.

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PLEISTOCENE GEOLOGY OF THE DOOR PENINSULA, WISCONSIN

BY F. T. THWAITES AND KENNETH BERTRAND

ABSTRACT

The area described completes the detailed mapping of the glacial geology of the Green Bay lobe. The drift belongs to two substages of the Wisconsin stage of glaciation. The major topographic features due to glaciation are the result of the Cary substage. Above this drift lies the till of a later substage, tentatively named Valders because its relation to the Mankato drift of Minnesota is unknown. Separating the two drifts is the Forest Bed which indicates a retreat of the ice margin far enough to free from ice both the Straits of Mackinac and the entire St. Lawrence valley. The advance of the Valders ice buried the two creeks Forest Bed with lake deposits formed in front of the ice. The age of this deposit by radiocarbon dating is about 11,400 years. The Valders till is red and is high in clay and silt because the ice readvanced over red lake clays formed during the Two Creeks interval of recession. Valders till smoothed and almost obliterated Cary land forms but was so thin over wide areas that it is no longer identifiable. These areas were not nunatacks. When the Valders ice receded, lakes were again formed in front of it. Four outlets of Lake Oshkosh in the Green Bay lowland across the Niagara cuesta to Lake Chicago in the Lake Michigan basin are described. Opening of the last outlet caused the two lakes to merge into Lake Algonquin. Ice was not far away then since varved clays accumulated in this lake. Further recession of the ice margin uncovered an outlet to the northeast which caused the water level to fall about 350 feet below that of the present. Evidence of this low level, now named Lake Chippewa, is described from this area. Rise of the land to the northeast caused the water to rise to the level of Lake Nipissing about 20 feet above the present. Erosion of the modern outlet of lake Huron caused the fall to present conditions.

CONTENTS

TEXT

	Page
Introduction	832
History of investigation	832
Location of area	832
Field methods	833
Previous surveys	834
Bedrock	835
General statement	835
Structure of bedrock	836
Topographic effects of bedrock	836
Pleistocene	838
Introduction	838
Form of the rock floor	838
Origin of basin of Lake Michigan	840
Glacial erosion of Niagara Escarpment	843
Caves and weathering in Niagara dolomite	843
Summary	843
Glacial deposits—tills	843
Composition	843
Color	844
Glacio-aqueous deposits	844
Glacio-fluvial deposits	844
Glacio-lacustrine deposits	845
Land forms caused by glacial deposition	845
Marginal deposits or end moraines	845
Ground moraine	847
Drumlins	847
Outwash	847
Eskers	847
Lake deposits	848
History of glaciation	848
Introduction	848
Pre-Cary glaciation	849
Pre-Cary interval	850
Cary glaciation	850
Direction of ice movement	850
Formation of end moraines	850
Sand ridges of Green Bay lowland	850
Formation of Kettle Interlobate Moraine	850
Lake-shore moraines	851
Moraines of Door County north of Sturgeon Bay	853
Soils of the Cary drift	853
De Pere-Greenleaf terrace	853
Early Lake Oshkosh	853
Early Lake Chicago	854
Two Creeks interval	855
Pro-Valders lake deposits	856
History of the Two Creeks interval	859
Age of the Forest Bed	859
Checks on age determination	862
Varve age determinations	862
Other age determinations	863

Page

Valders glaciation ... 864
Nomenclature ... 864
Lobation and direction of flow of Valders ice ... 864
Nature and extent of Valders till ... 865
Soils on Valders till ... 865
Erosion by Valders ice ... 866
Drainage of Valders ice ... 866
Lake Shoto ... 867
Later Lake Oshkosh ... 867
Lake Manitowoc ... 867
Outlets ... 867
Drainage of later Lake Oshkosh ... 868
Later Lake Chicago ... 868
Lake Algonquin ... 868
Low level of Taylor lakes ... 869
Shore-line delineation ... 872
Cause of rise in water level ... 872
Correlation of the low stage ... 873
Uplift of the Nipissing beach ... 874
Cause of uplift ... 874
Postglacial weathering ... 875
Postglacial stream erosion ... 875
Postglacial wave erosion ... 875
Postglacial wave deposition ... 876
References cited ... 877

ILLUSTRATIONS

Figure Page

1. Location of area ... 833
2. Cross sections showing relation of drift to bedrock surface ... 837
3. Contour map of bedrock surface between Kaukauna and Brillion ... 839
4. Contour map of bottom of Lake Michigan ... 841
5. Cross sections through the basin of Lake Michigan ... 842
6. Mechanical composition of Valders and Cary tills of northeastern Wisconsin and part of southern Michigan ... 846
7. Sketch of face of quarry at Valders, Wisconsin ... 851
8. Map and cross sections showing Cary gravel deposits interpreted as deltas deposited in early Lake Chicago ... 852
9. Contour map of eroded terrace near Green Bay ... 854
10. Contour map of eroded terrace near De Pere ... 855
11. Summer maximum levels of Lake Michigan from 1873 to 1953 ... 857
12. Exposures of the Two Creeks Forest Bed and associated deposits ... 857
13. Changes of water levels in Lake Michigan basin ... 860
14. Cross sections at Sturgeon Bay, Kewaunee, and Two Rivers ... 870
15. Map of Lake Chippewa, the low stage in the Lake Michigan Basin between the Algonquin and Nipissing stages ... 871
16. Profiles of the warped beaches of the Door Peninsula ... 873
17. Sketch map showing shore-line recession and stream intercision ... 876
18. Section along shore of Lake Michigan near Two Rivers ... 876

Plate Following page

1. Glacial-erosion topography and buried end moraine ... 844
2. Varved clays and drumlins ... 844
3. Overridden Cary moraine and varved clays ... 844
4. Two Creeks Forest bed ... 844
5. Two Creeks Forest bed ... 844
6. Contacts of Valders till with pro-Valders lake sediments ... 858
7. Pro-Valders deposits near Green Bay ... 858

Facing page

8. Glacial geologic map of area ... 880

TABLE

Table Page

1. Radiocarbon (C^{14}) determinations from Forest Bed ... 861
2. Radiocarbon Ages of Stages in later History of Lake Michigan ... 874

Introduction

History of Investigation

In 1922 Thwaites spent several weeks in Door County and thence south to Two Rivers. He recorded many glacial striae and investigated the bedrock geology for oil possibilities. The geology was mapped in detail near Sturgeon Bay and to the northeast, a region of very thin drift cover. L. R. Wilson studied the Two Creeks Forest Bed intensively in 1932, 1935, and 1936 under the direction of Thwaites, and in 1935–1937 Bertrand surveyed the human geography of the entire area of red glacial till and recorded some data on glacial features. From 1925 to 1956 the Forest Bed has been visited annually (with the exception of 1942–1945) by classes of students from the University of Wisconsin, by one of the excursions of the International Geological Congress in 1933, and by many other groups and individuals. Starting in 1948 the authors have visited the area to determine elevations, examine critical areas, collect samples of till and wood, and take photographs.

Location of Area

Figure 1 shows the position of the area here discussed relative to those surveyed by Alden to the south and by Thwaites to the west. Although the present report is not confined to the Door Peninsula, whose southern limit extends east from Green Bay, the name is used to avoid

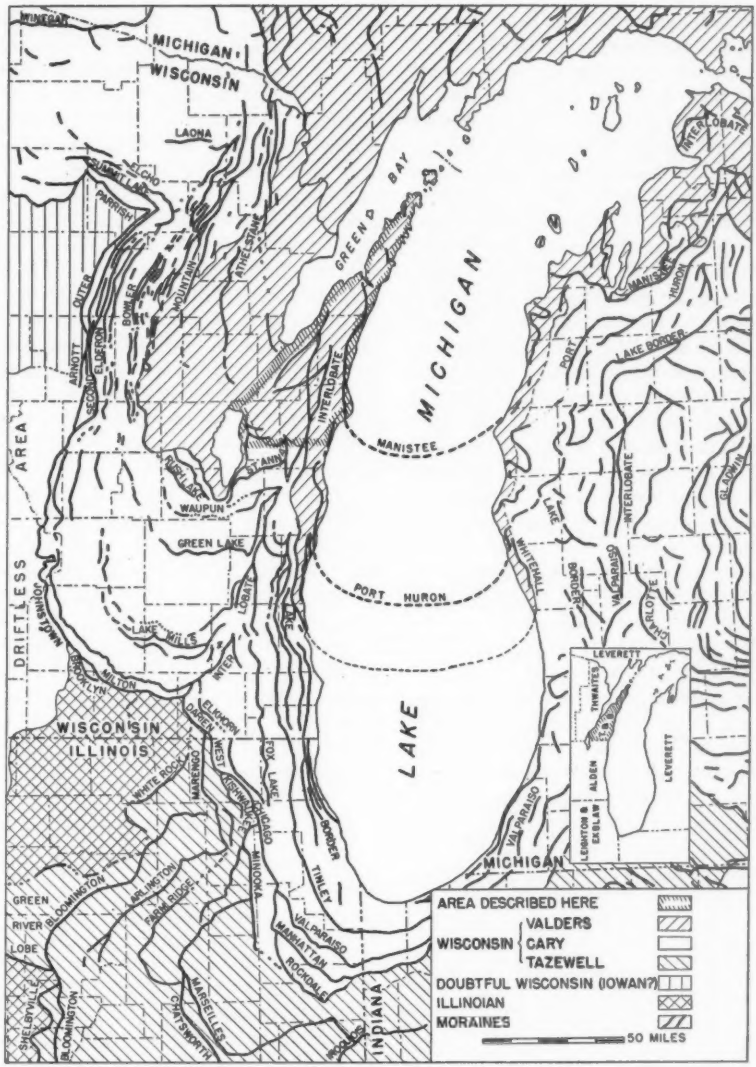

FIGURE 1.—LOCATION OF AREA

Showing relation to drifts and end moraines. Based on work of Leverett, Alden, Leighton, Ekblaw, and Thwaites, with some modifications. Farmdale drift omitted.

confusion with the older report by Thwaites (1943) on part of northeastern Wisconsin. There is a narrow overlap onto Alden's area (1918) to the south because the south limit of this map is fixed by township lines and not by a parallel of latitude.

Field Methods

At the start of the survey the only available topographic maps of the Door Peninsula were limited to the Neenah quadrangle of the U. S. Geological Survey and to a narrow band along

some of the lake shores and Fox River surveyed by the U. S. Lake Survey; hence attention was directed to finding elevations on which to base a contour map, details of which were filled in by use of air photography. A sensitive Wallace and Tiernan barometer was loaned by the Water Resources Branch of the United States Geological Survey at Madison, thanks for which courtesy is here expressed. Check elevations were obtained from railroad profiles, U. S. Coast and Geodetic Survey bench marks, and highway surveys. The air photographs were purchased by the University of Wisconsin as Project RF 47:312 for which our thanks are also expressed. A later grant R. F. 54:295 and a gift of $50.00 from Bruno Hartman, Consulting Engineer of Sheboygan, were both devoted to drafting, and thanks are expressed for both of them. Because the flights on which these photographs were taken were north-south on alternate section lines, the plan was to find the elevations of every section corner or road intersection on the section lines common to two flights, with such check points near the corners of every stereo-pair. This was not everywhere possible because of local irregularities in the mostly rectangular road pattern and other features of the terrain, including lack of ground points which could be definitely located on the photographs. Since most of the railroads run across the area in a general east-west direction, elevation checks were available at rather short travel-time intervals. Some checks taken on lake elevations proved satisfactory in the case of Lake Michigan but less so on Green Bay because there is not the consistent onshore wind which comes from the larger body of water in summer. Some east-west check lines were also run with the barometer. Repeated determinations of the same point were very satisfactory. Almost all the highway levels and U. S. Geological Survey telescopic-alidade points obtained after the original survey agreed very closely. Most of our elevations are correct within 5 feet, and a discrepancy of more than 10 feet is almost invariably an error either in recording a location or in reading the scale. However, in drawing correction curves it is obvious that on many days, especially in hot, windy weather, changes in atmospheric pressure are a succession of "waves" with an amplitude of 10 feet or more. The time position of a check reading on a known elevation with regard to these irregularities can well cause a number of successive readings to be consistently in error. Since 1951 the U. S. Geological Survey has been surveying a number of quadrangles east and southeast of Green Bay. Advance copies of these quadrangles were available and could be compared with the map previously made by the authors. Their contouring with the Kelsh plotter agrees closely with ours which was done with a Fairchild stereocomparagraph. Important trouble due to tilt of the photographs was almost entirely confined to those near the shores of Lake Michigan and Green Bay. In very flat areas there is much difficulty in drawing contours from air photographs, and plane-table surveying by the U. S. Geological Survey delayed publication of the quadrangles.[1]

A serious difficulty in mapping Pleistocene land forms in the region surveyed is due to the fact that all but a few such land forms are older than the last ice advance. The cover of red glacial till smoothed and almost buried many of the important land forms which were the product of an older glaciation. Almost all boundaries, such as those of drumlins and end moraines, were determined by stereoscopic examination of air photographs. Borders of outwash of the last ice advance are based almost entirely on soil maps.

Previous Surveys

The first published map of the glacial deposits of the Door Peninsula is that by T. C. Chamberlin (1877) who surveyed the area for the Wisconsin Geological Survey. Chamberlin, like several other early geologists (Winchell, 1871), set up a stratigraphic column for the Pleistocene deposits as he did for the bedrock formations. On his small-scale geologic map in the atlas, the only topographic feature shown is the Kettle Moraine. Chamberlin concluded that the red glacial drift was a lacustrine deposit and divided it into upper and lower divi-

[1] The 15-minute quadrangles published up to 1956 include: Green Bay, New Franken, Casco, DePere, Denmark, Kewaunee, Neenah, Chilton, Reedsville, Manitowoc, Kiel, Howards Grove, Sheboygan North, 7½ minute quadrangles include: Green Bay West, Green Bay East, DePere, Bellevue, Manitowoc, School Hill, St. Wendell, Cleveland.

sions separated by beach deposits. He accounted for the variation in elevation of the margin of the "lower red clay" by subsequent earth movement. In 1905 J. W. Goldthwait (1907) studied the abandoned shore lines of the district and discovered the Forest Bed at Two Creeks. In 1918 the final report of Alden's studies of the Pleistocene south of 44° N. Lat. appeared. Alden presented proof that the "red clay" is a glacial till formed by plowing up of older lake clays and that this deposit is older than the beaches measured by Goldthwait. Although the northern limit of Alden's map is approximately the south limit of this report, he did considerable work north of that line, the results of which are presented on Plate XXXVI of his report. Copies of Alden's field notes and field maps are on file in the office of the Wisconsin Geological Survey and aided materially in preparing the present map (Pl. 8). Alden did not visit the Forest Bed, and it was not until 1930 that Wilson began his intensive study of that deposit although parts of it farther west had been previously described by Lawson (1902). Maps of portions of Door County published by R. R. Shrock (1940) and O. L. Kowalke (1938; 1946; 1952) were used in the preparation of the present report. Murray (1953) published results of mechanical analyses of some of the till samples collected by the authors. The 1956 mimeographed guide book of the "Friends of the Pleistocene" in lower Michigan, was also used as a source of information.

Bedrock

General Statement

The bedrock formations of the Door Peninsula and adjacent parts of northeastern Wisconsin are marine sediments of Ordovician and Silurian age. Knowledge of them was obtained from the work of Chamberlin (1877), Shrock (1939; 1940), Karges, (1936), Buckley (1898), Steidtmann (1924), Savage (1916), Hotchkiss and Steidtmann (1914), and Thwaites (1913; 1923; 1931). Although outcrops are numerous along the escarpment in the west and northwest parts of the area and in northern Door County, it is difficult to ascertain thicknesses of strata except from information derived from drilling of deep wells. The following column briefly summarizes existing information on the exposed rock formations.

Silurian

Niagara formation
- Racine: dolomite, light gray, massive to thin-bedded, dense to granular texture, many bioherms, more than 400 feet
- Manistique-Cordell: dolomite, light gray, thin-bedded, cherty, fossiliferous, many bioherms (Upper Coral of old reports), about 90 feet
- Schoolcraft: dolomite, light gray, thin-bedded, fossiliferous, (Lower Coral of old reports), about 70 feet
- Burnt Bluff-Hendricks: dolomite, light gray, thin-bedded, very fine-grained to granular texture (generally included with Lower Coral in older reports), about 30 feet
- Byron: dolomite, light gray, thin-bedded, mud-cracked and ripple-marked, only slightly fossiliferous, few bioherms, about 100 feet
- Mayville (Alexandrian): dolomite, light gray, cherty, thick-bedded, fossiliferous, about 100 feet

Greatest known thickness of Silurian, 735 feet

Ordovician

- Neda: hematite, oölitic, red, calcareous, occurs in lenticular, isolated deposits, known only near Little Sturgeon Bay, De Pere, and in wells at Manitowoc and St. John, maximum 55 feet
- Richmond (Maquoketa): shale, blue gray, some brown, dolomitic; dolomite, light gray to blue gray, shaly, thin-bedded, occurring in many lenses, and especially at top in Door County, less than 300 to more than 500 feet
- Galena-Platteville: dolomite, light gray to blue gray, thick-bedded above to thin-bedded below; exposed only along Fox River, about 200 feet

Because the younger formations are most widespread in the area this column is discussed from top down.

The Silurian dolomites, including the Alexandrian or Mayville dolomite at the base, are commonly mapped as a unit in Wisconsin and are termed the Niagara formation. It forms the cap rock of the Niagara cuesta which makes up the ridge between the basin of Lake Michigan and the Green Bay-Lake Winnebago lowland. Attempts to subdivide this dolomite formation into either formations or members by using insoluble residues, grain size, or fossils have not been uniformly successful because of the presence of many reefs or bioherms. Adjacent to these organic accumulations, the strata dip at rather steep angles. It is difficult to discriminate between such original dips and inclinations due to earth movement. Most of the Niagara dolomite is pure. It was formerly

quarried for lime at many places within the area (Bertrand, 1946). Exhaustion of the cheap fuel supply from the northern forests plus changes in the building industry have closed most of the quarries. This fact makes the search for glacial striae now much more difficult, for little fresh stripping is available.

The oölitic iron ore of the Neda is very limited in distribution and pinches out in a distance of a few feet as can be observed at the type locality near Mayville, Wisconsin (Thwaites, 1913). Exploration within this area has been negligible, for the ore is low grade, very friable, and high in phosphorus. The best-known outcrop is at Katells Falls in SE ¼ NE ¼ sec. 32, T. 23 N., R. 21 E.

The Richmond or Maquoketa shale is exposed in many localities along the escarpment from the Calumet County Park near Stockbridge to Sherwood Point at the entrance to Sturgeon Bay where it dips below the level of Green Bay. Attempts to subdivide the formation on a lithologic basis have not been successful. However, dolomite near the top seems to increase toward the northeast. A well in Peninsula State Park (Shanty Bay camp ground) penetrated more than 300 feet of shaly dolomite below the base of the Mayville. The correlation of these strata with the section in Michigan is not here attempted. The base of the shale, which rests with abrupt contact on the underlying Galena dolomite, is nowhere exposed.

The Galena dolomite is exposed only along Fox River which is the northwest border of the map (Pl. 8). The river-bed outcrops of this formation caused the rapids in Fox River which in turn brought about the location of the paper and pulp industry which is so important along that stream.

Structure of Bedrock

The prevailing structure of the bedrock formations is a uniform dip to the south of east at an angle of less than 1°. A few faults are known. One is indicated by a discrepancy between the sections exposed in two adjacent quarries just north of the south part of Sturgeon Bay. It appears to trend south of west from there and may be the cause of the inclined bedding on the south side of a little creek in center sec. 33, T. 27 N., R. 24 E about 2½ miles northeast of Brussels. Displacement is down on the southeast side. Records of two wells at Algoma about 200 feet apart indicate another fault with the same strike and displacement. Poor surface exposures near Greenleaf may indicate another fault. Thwaites (1931) formerly mapped faults along both sides of Rawley Point, north of Two Rivers. This conclusion was based on a driller's log of an old well in Two Rivers which does not check with newer wells near by. It is now concluded that no such faults exist in the vicinity of Two Rivers.

Topographic Effects of Bedrock

The major feature of the area is the Niagara Cuesta which forms the high escarpment east of Lake Winnebago and Green Bay, the backbone of the Door Peninsula. The highest bedrock in this ridge rises more than 1500 feet above the bottom of Lake Michigan to the east. The cause of the escarpment is the marked difference in resistance to erosion between the thick-bedded Niagara dolomite and the underlying Richmond shale and shaly, thin-bedded dolomite. The escarpment is not everywhere conspicuous because of drift cover and has marked re-entrants. The Mayville dolomite caps the escarpment from the south border of the area north to the entrance to Sturgeon Bay. At the top of this member (or formation) there is a bench or terrace which in many places is miles wide. North of Sturgeon Bay the base of the Mayville is concealed below the water level. Only small areas of the Mayville bench occur there, and these terminate at Peninsula State Park. The outcropping margin of the thin-bedded Byron rises from the Mayville bench in a well-marked escarpment about 100 feet in height which can be traced from Washington Island south through almost all of Door County. The outline of this escarpment is irregular, and there are a number of isolated outliers. It is very conspicuous in the landscape, especially at Pottawatami State Park near Sturgeon Bay and south of Brussels. The strata that overlie the Byron do not form prominent escarpments. Throughout the dip slope east of the Byron escarpment there are many isolated rock hills. Several of these hills were opened by quarries, for instance at Grimms, Quarry, and Valders

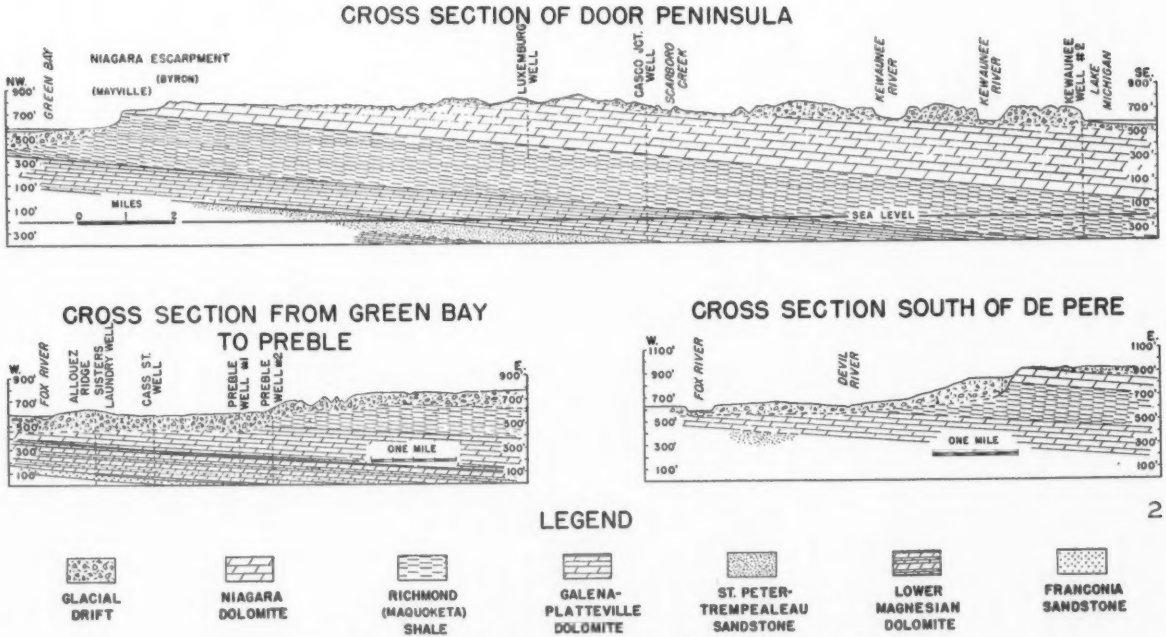

FIGURE 2.—CROSS SECTIONS SHOWING RELATION OF DRIFT TO BEDROCK SURFACE

in Manitowoc County, and it is certain that their resistance to erosion is due to a reef or bioherm core over which the strata arch. Hills not thus exposed but which appear from their shape to be probable counterparts include, among many others, the "North Bay dome" at the southeast corner sec. 34, T. 31 N., R. 28 E., Stevens Hill at Sturgeon Bay, a hill in secs. 22 and 23, T. 25 N., R. 23 E., in northwest Kewaunee County, several hills near by, and Oak Hill, the highest point in the area (elevation about 1140 feet above sea level).

Pleistocene

Introduction

The present study of the Pleistocene deposits of the Door Peninsula and adjacent territory is among the first to use air photographs for the mapping of land forms. Considerable attention was also directed to the mechanical constitution of the deposits themselves, particularly the till. The distinct difference in physical makeup of the two tills exposed in the region also permits an approach from the stratigraphic standpoint, a point of view decidedly reminiscent of that employed by some of the earlier students of the district. The organic deposits known as the Forest Bed which occur between the two tills are also important in this method of study, which depends more upon laboratory determinations than upon direct tracing of boundaries in the field.

Form of The Rock Floor

North of Sturgeon Bay rock outcrops are numerous and extensive areas are covered with only a few inches of till. Drift more than 20 feet thick is exceptional. To the south, exposures of bedrock are decidedly less abundant, and drift locally reaches 450 feet thick. The Niagara escarpment is very abrupt; at its foot occurs the thickest drift. The lowest bedrock surface known at Green Bay is at Liebmann Packing Company Northeast of the city, elevation 425 feet above sea level. West from this point the rock surface rises steadily and crops out near Duck Creek, west of the area here considered. High drift bluffs occur at several points along the shore of Lake Michigan, although deep drift-filled valleys are unknown. In the Sturgeon Bay ship canal bedrock is found at a depth of about 20 feet, and in the Bay to the north the depth of drift (Fig. 13) does not exceed 84 feet below the surface of the water.

The area around Forest Junction in the southwest part of the area (Fig. 3) is the major exception to the rule of relatively thin drift. This area was mapped in great detail by means of numerous well records supplied by the late J. J. Faust, well driller of Kaukauna.

Driller's Log of well in NW¼NW¼ sec. 27, T. 20 N., R. 20 E.

J. J. Faust and Sons, Drillers

	Thickness, feet	*Depth, feet*
Clay	90	90
Sand	21	111
Clay	115	226
Muck and sand	8	234
Clay	216	450
Shale	10	460
Limestone	6	466

Driller's Log of Peter Reuther well in N. part sec. 28, T. 20 N. R. 20 E.

J. J. Faust and Sons, Drillers

	Thickness, feet	*Depth, feet*
Clay	100	100
Rotten wood, moss, snail shells, gas	4	104
Sand, fine	30	134
Clay, red, some white streaks, a few stones	238	372
Shale	6	378

Although the deepest drift proved by samples is only 115 feet (in the village wells at Brillion), there seems little doubt that these driller's logs actually show drift material to the depths indicated. Figure 3 shows that this deep-drift area is almost surrounded by thin drift, proved by many outcrops. It is very similar to the area of deep drift at Black Creek described by Thwaites (1943, p. 116–119). It is difficult to prove that such a deep depression could be due to glacial erosion, for there is no evident reason why this process was localized here. The Richmond shale extends far from the Niagara escarpment and forms a large hill in the southwest corner of Brown County near Holland, the "shale ledge." If this depression were due to glacial erosion, it would be necessary to assume that the ice was moving south or southwest

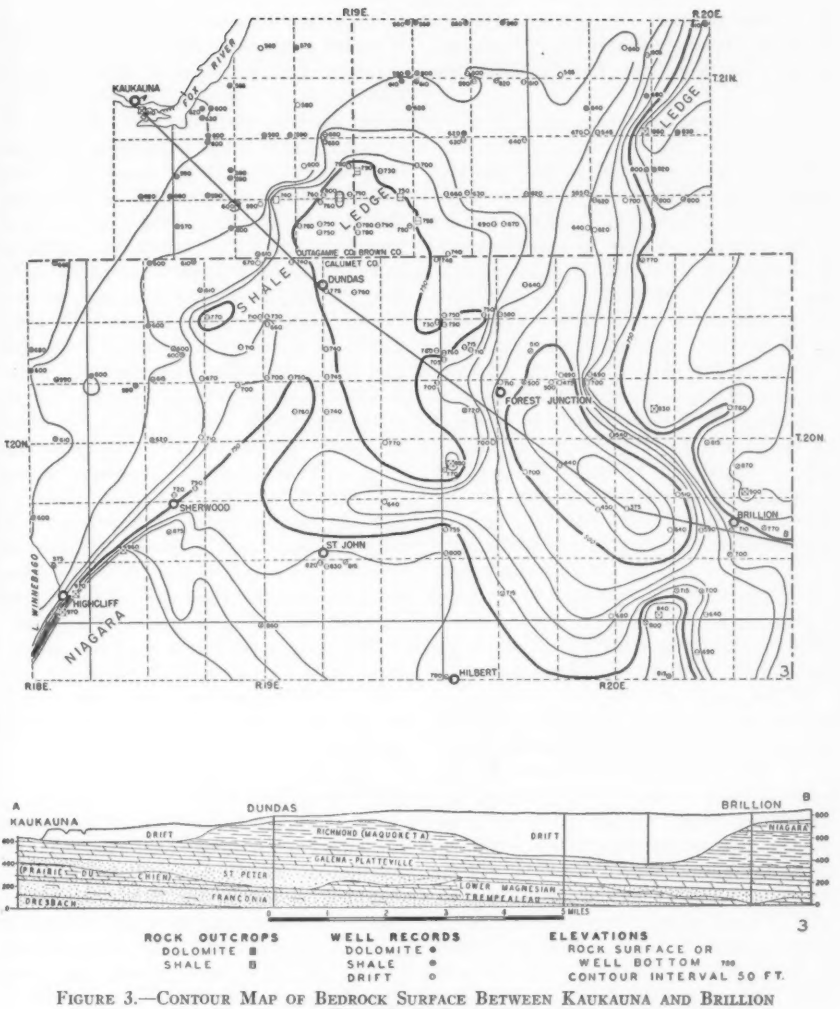

FIGURE 3.—CONTOUR MAP OF BEDROCK SURFACE BETWEEN KAUKAUNA AND BRILLION WITH CROSS SECTION

descending the escarpment by extrusion flow. A southwesterly direction of an early ice advance is in fact suggested by some striae, including some reported in field notes by Shrock at Brillion, which trend 245° (S. 65° W.).[2] This area of thick drift apparently is enclosed, and hence it is difficult to regard it as the head of a deep valley which discharged eastward into the Lake Michigan basin. It has also been suggested that it might be the plunge pool of a glacial river which crossed the Niagara escarpment at this place. If this were the case, the outlet would have to be north along the face of the escarpment, but there is no apparent line of escape to the north or west. On the whole, this idea of a plunge pool appears even less likely than that of glacial erosion by ice

[2] Throughout this paper all directions are given in azimuth, that is degrees to the right of north around the horizon.

either from the northeast or the northwest. The Brillion basin apparently was formed before the Cary substage of the Wisconsin glaciation.

Origin of Basin of Lake Michigan

The origin of the basin (Figs. 4, 5) that now contains Lake Michigan was long controversial. Early geologists lacked information, especially on the subsurface bedrock geology. Owing to the easterly dip of the bedrock formations the rocks that underlie the basin of Lake Michigan can be studied only in drill holes. Wildcatting for oil has now supplied much information which the writers obtained from the Michigan Geological Survey. Furthermore, the pioneers considered only a single process of origin of the lake basin. From the study of modern charts of the lake and of many well records, it is obvious that several processes must be considered including: (1) obstruction of preglacial valleys by drift deposits, (2) preglacial erosion and solution of bedrock, (3) earth movement, and (4) glacial erosion. In regard to (1) it has never been proved that there is more than about 250 feet of drift at the south end of Lake Michigan, and the maximum known depth of water is 984 feet, which extends slightly more than 400 feet below sea level (Fig. 4). The cross sections of Figure 5 show that the deep basin east of Kewaunee is underlain by the salt- and gypsum-bearing strata of late Silurian and Early Devonian age. The amount of drift cover over the sublake strata is unknown. Studies by Landes and others (1945) of the vicinity of the Straits of Mackinac demonstrated that solution of these evaporite formations was extensive in pre-middle Devonian time and that it resulted in extensive brecciation of the overlying strata. Solution is independent of base level, for the depth to which fresh water can penetrate is governed by the relief of the water table which supplies the necessary head or energy for movement. Fresh underground waters extend to considerable depths causing solution where there is, first, adequate pressure difference to overcome friction and, second, no shallower avenue of flow that can take care of the discharge. The thin-bedded dolomites associated with the evaporites would also be readily eroded by surface runoff. Preglacial lowlands were undoubtedly eroded both on the Ordovician shales of pre-Niagaran age and on the post-Niagaran evaporites. Martin (1916, p. 271) suggested that the valley in which Sturgeon Bay now lies is a surviving segment of Menominee River which flowed down dip into the Michigan basin. If this interpretation is valid, the preglacial level of the evaporite lowland probably was not more than 100—200 feet lower than the present surface of the lake. It has long been known that the region has undergone postglacial and recent uplift toward the northeast. The bottom of Lake Michigan, which is crossed by an escarpment of middle Devonian limestones extending northeast from Sheboygan, Wisconsin, is not in harmony with either the attitude of the glacial lake beaches (Fig. 16) or the present-day changes of level described by Moore (1948). Although earth movement is an undeniable factor, it is glacial erosion that explains not only the great depth but also the irregularity of the bottom of the Lake Michigan basin. Shepard (1937) stressed these facts in support of the idea of ice erosion, but at the time he wrote did not know of the extensive brecciation of much of the bedrock of the northern portion of the basin and adjacent parts of Lake Huron (Landes and others, 1945). The brecciated limestones, some of which are no more resistant than unconsolidated gravel, would be much more easily eroded by ice than the solid, thick-bedded dolomites of the Niagara. Such erosion would be aided by glacial plucking and grinding of thin-bedded strata plus new solution of the salt by the ice and its meltwater. Glacial erosion is a mechanical process, governed by the physical nature of the underlying rock and not by its chemical composition. If this rock character permits ice to intrude between and freeze around fragments of some size, erosion is necessarily much faster than by grinding the bedrock into powder. It is more difficult to appraise the effect of ice coming into contact with salt, for, although the salt beds are mechanically weak, they would tend to melt the ice which touched them. The irregular topography of the north part of the Lake Michigan basin extends far inland into southern Michigan and explains not only many of the islands in Lake Michigan but also the large kettle lakes of the northern part of the Lower Peninsula. It may well be that much of the solution was by ground

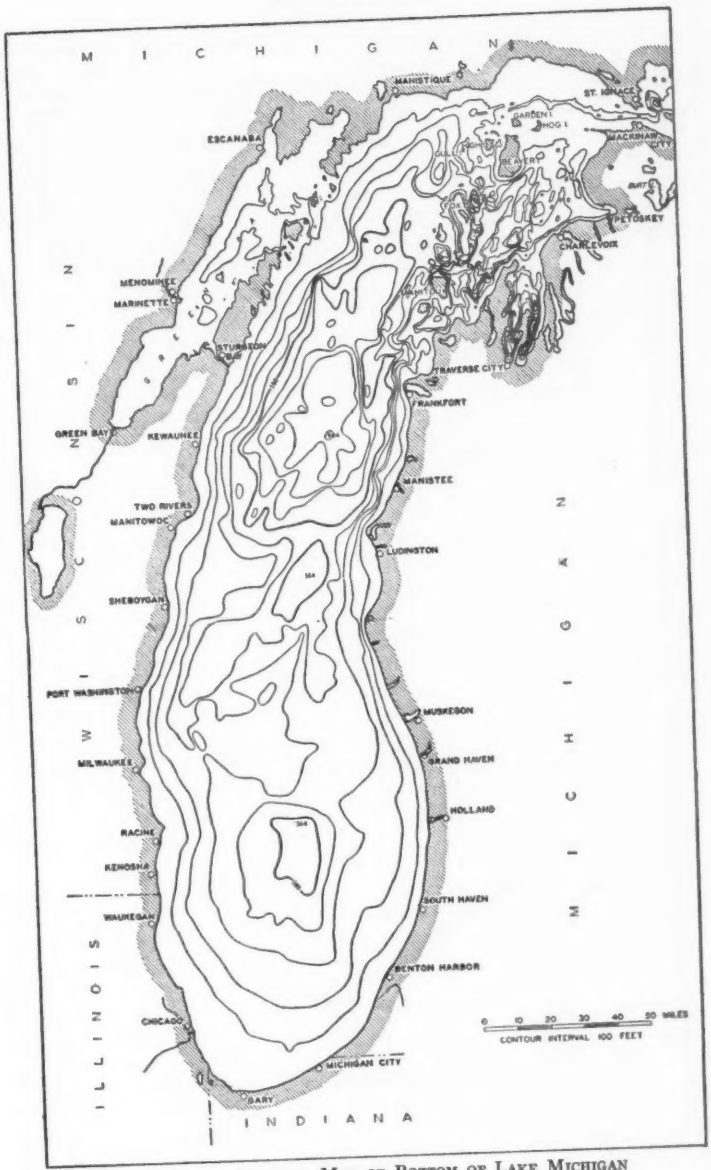

FIGURE 4.—CONTOUR MAP OF BOTTOM OF LAKE MICHIGAN
Based on U. S. Lake Survey

water in post-Dundee-Traverse time. The contrast of this irregular bottom topography with the broad smooth depression on the Devonian shales opposite the south line of Wisconsin and between the Traverse and Dundee escarpments farther north is very striking. The limestone cuestas are broken through by a valley parallel to the east side of Lake Michigan, but there is

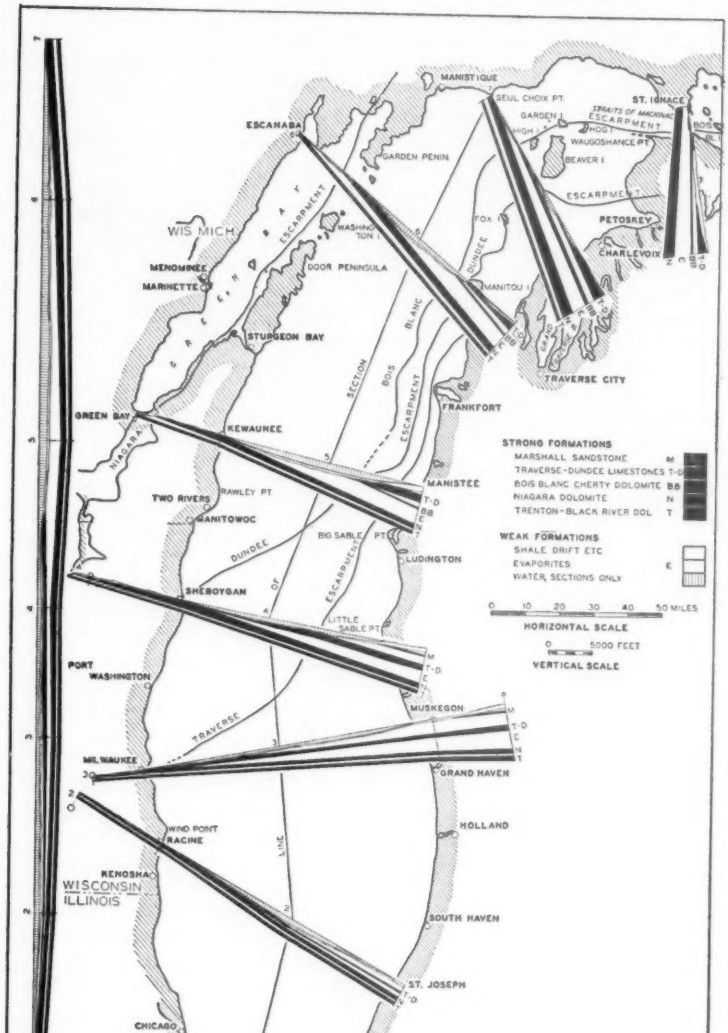

FIGURE 5.—CROSS SECTIONS THROUGH THE BASIN OF LAKE MICHIGAN
Based on well data

no suggestion of its continuation to the south into northern Indiana, which should be present were it a relic of a preglacial line of drainage. Its presence in no way vitiates the theory of glacial erosion as the predominant cause of the basin. Hough (1955) calls this the valley of "Grand Haven River" which flowed at the time of the low level of Lake Michigan (Fig. 14).

Glacial Erosion of Niagara Escarpment

The escarpment at the edge of the Niagara dolomite presents two phases: (1) straight-cliffed sections, and (2) less uniform drift-covered portions. As shown in Figure 1 of Plate 1, some of these cliffs are as straight as if built by man. Starting at the south: (1) the escarpment near High Cliff trends 34° for 3 miles. (2) a portion near Greenleaf trends 22° for 2 miles, (3) for 5 miles south of DePere it bears 33°, (4) for 6 miles near Bay Settlement it bears 24°, and (5) for 3 miles north of Brussels it bears 36°. Martin (1916, p. 230–235) used this phenomenon as evidence of important glacial erosion. He also stressed the absence of small outliers of the Niagara dolomite. This conclusion was apparently correct. The straight portions are undoubtedly related to glacial plucking controlled by joints. However, it is difficult to check this. Adjacent to quarries and road cuts blasting has so broken the rock that joints occur in almost all directions. The major or master joints are obscured, except in old workings, where sweet clover and weeds spring up in them so that the trend is easily determined. At Greenleaf the most abundant joints bear 18°–25° thus bracketing the 22-degree direction of the escarpment. Farther north, where the cliff bearing averages 33°, joints are abundant at 20° and from 35° to 42°. Although no measurements were made at the other localities these data substantiate the hypothesis.

Caves and Weathering in Niagara Dolomite

Martin (1916, p. 236–238) also employed the lack of residual clays and rarity of caverns as evidences of glacial erosion. However, he did not know of either the Egg Harbor (SE ¼ NW ¼ sec. 3, T. 29 N., R. 26 E.) or Maribel (center sec. 13, T. 21 N., R. 22 E.) caves, nor did he realize that clay derived from the weathering of the dolomite is found in several places on the Door Peninsula. Weathered joints and bedding planes have been distinguished in deep wells far below the present surface by discoloration of the dolomite. The Niagara is for the most part pure dolomite which would leave little residium and which is not so readily soluble as are calcitic limestones; caverns are rare in the unglaciated Niagara dolomite of northwestern Illinois. One of the best evidences of extensive glacial erosion is the abundance in the glacial drift of pebbles, cobbles, and boulders derived from the Niagara dolomite. Boulders up to 10 feet in length are not uncommon and must have been plucked from near-by bedrock. At High Cliff a large mass of Niagara dolomite is now detached from the cliff by a crevice several feet across. It is tilted toward the valley and must have moved in postglacial time. It is mute evidence of why the glaciers removed blocks of dolomite from above the wet, slippery shale below and incorporated them into the drift.

Summary

Despite the local occurrence of highly weathered bedrock and of a few caverns, glacial erosion must certainly be regarded as one of the most potent forces in shaping the bedrock which controls the present landscape. No data on the preglacial level of the crest of the Niagara escarpment are available. The production of a Byron escarpment on the thin-bedded dolomites fits better the hypothesis of glacial erosion than of normal weathering and stream erosion, and this is corroborated by the great amount of fresh Niagara debris in the drift, the irregularity and depth of the bed of Lake Michigan, the discordance of level between the bottom of the big lake and that of Green Bay, the lack of any evidence of a graded rock valley which could have once drained the basin of Lake Michigan, and the regularity of the escarpment. The caverns may have been formed below the water table at some ancient time and on account of their rarity and difficulty of discovery have little importance as evidence.

Glacial Deposits—Tills

Composition

The glacial tills of the Door Peninsula vary widely in physical composition (Fig. 6). Adjacent to the Niagara bedrock the till is a rubble of subangular fragments nearly all of local derivation. A count of 117 boulders more then 6 inches in diameter taken in NE ¼ NE ¼ sec. 15, T. 22 N., R. 23 E., gave Niagara dolomite 73 per cent, light-colored igneous rocks 8 per cent, dark-colored igneous rocks 16 per cent, and miscellaneous rocks 3 per cent.

Because igneous rocks break into larger fragments than do most dolomites, the finer stones of the till should show a higher percentage of the local dolomite. Studies on comparatively large rock fragments do not, however, give as clear a picture of the nature of the till as do mechanical analyses of the finer portions. However, complete analyses, such as are made for soil studies, are unnecessary for this purpose, and the graphs of Figure 6 show only pebbles and granules more than 2 mm in diameter, sand from 2 mm to ⅛ mm, and very fine sand, silt, and clay all smaller than ⅛ mm. The results demonstrate the wide size variability, which is characteristic of all tills, as well as the influence of near-by sources of material. Underlying sand is reflected in a high sand fraction. Those tills near bedrock show the highest percentage of the larger particles, and those underlain by lake clays show a very high proportion of the small sizes. Presence of lake clays deep in the basin of Lake Michigan cannot be confirmed from present information but seems probable. Tills near the shore show a high silt-clay content. The carbonate content of the till has been determined by Robinson (Unpublished) and by Murray (1953). The former used all sizes of fragments and concluded that it exceeds 25 per cent expressed as $CaCO_3$. Murray confined his determinations to particles less than 4 mm diameter and used a different technique. He arrived at the figure of 7 per cent for the gray tills and slightly less than 10 per cent for the red tills. His x-ray examination of the clay-sized particles disclosed more dolomite in the gray tills than in the red tills.

Color

The tills of northeastern Wisconsin can be divided into two great groups on the basis of color: red and gray. In many exposures the distinction is as definite as in a colored diagram. The red till is everywhere found above the gray till. As determined by use of the Munsell color charts the red tills are classed as pale red when dry (5 YR 6/3) and moderate red when moist (5 YR 4/3 to 5 YR 5/2). The red color is due primarily to the content of red lake clays. This is evident even where the clay is much diluted with sand. The gray tills vary more in color from pale brown to pinkish gray (5 YR 6/2 to 10 YR 6/3) when dry to dark brown or dark reddish gray (7.5 YR 5/2 to 5 YR 4/2) when wet. These colors were checked by F. D. Hole, Associate Professor of Soils.

Glacio-Aqueous Deposits

Glacio-Fluvial Deposits

The meltwaters of the glaciers deposited a vast amount of sand and gravel within the area mapped. In Door County the areas mapped as moraines are predominantly very coarse, poorly

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Plate 1.—GLACIAL-EROSION TOPOGRAPHY AND BURIED END MORAINE

Figure 1.—Glacially eroded Niagara Escarpment. Stereopair of air photographs by Agricultural Adjustment Administration, 1938, showing a very straight portion of the Niagara Escarpment south of De Pere. The cause ascribed is glacial erosion controlled by plucking along joints.

Figure 2.—Cary end moraine buried by Valders till. Stereopair of air photographs by Agricultural Adjustment Administration, 1938. The mottled appearance, which is more prominent with some crops than with others, is due to relief of 10 to 15 feet, in part accentuated by gulleying. Settling of the Valders till over an irregular end moraine surface of Cary till is believed to be the cause.

Plate 2.—VARVED CLAYS AND DRUMLINS

Figure 1.—Varved clays disturbed by Valders ice. The varved clays deposited on top of the Cary till, which forms the lower part of the exposure, were disturbed by the Valders ice which left very little till at this locality. Clay pit in NW ¼ SE ¼ sec. 24, T. 19 N., R. 23 E., Manitowoc, Wisconsin. Vertical marks were made by a power shovel.

Figure 2.—Drumlins. Air photograph by Agricultural Adjustment Administration, 1938, showing drumlins south of Sturgeon Bay. The material is Cary till because the Valders till was so thin here that it has been all affected by soil-making forces. Drumlin areas are readily identified in air photographs by the effect of their form on field boundaries.

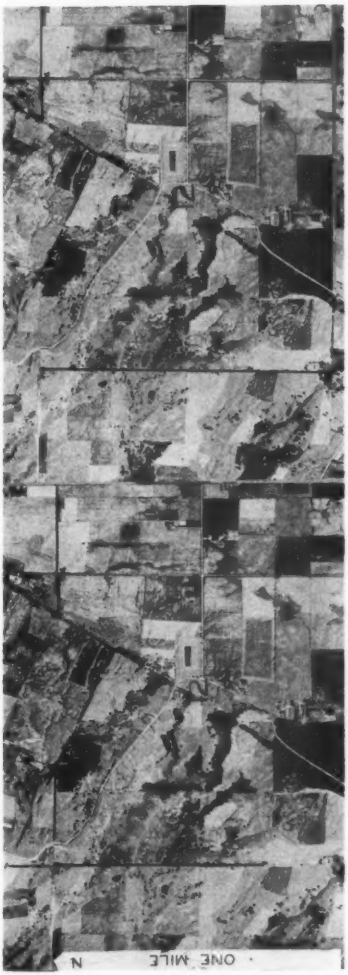

FIGURE 1

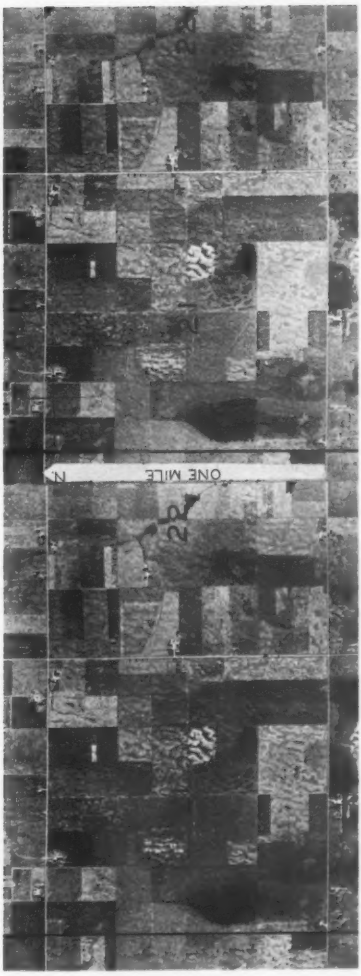

FIGURE 2

GLACIAL-EROSION TOPOGRAPHY AND BURIED END MORAINE

FIGURE 2

FIGURE 1

VARVED CLAYS AND DRUMLINS

FIGURE 1

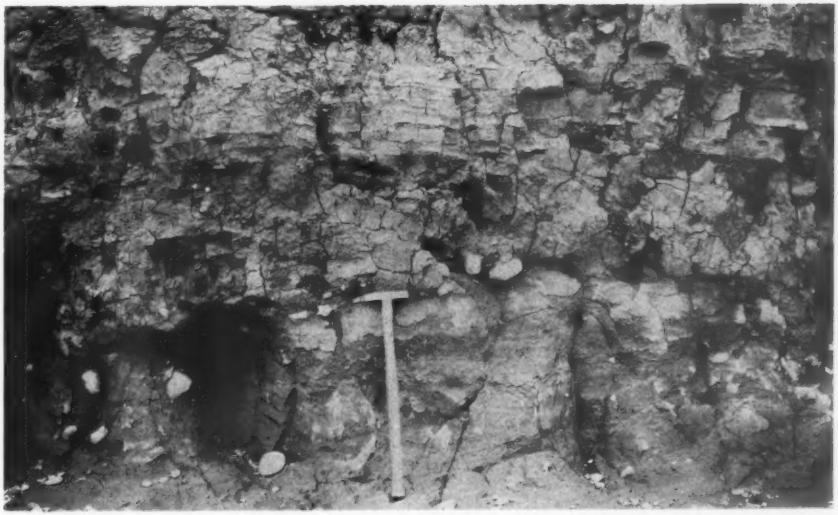

FIGURE 2

OVERRIDDEN CARY MORAINE AND VARVED CLAYS

TWO CREEKS FOREST BED

FIGURE 2

FIGURE 1

sorted gravel. Farther south, the Interlobate Moraine of Kewaunee and Manitowoc counties is to a large extent made up of gravel except that in many localities there is a thin cover of red glacial till. East and southeast of Green Bay the Niagara escarpment is fronted by a terrace of sand and gravel (Figs. 2, 9, 10; Pl. 1, fig. 1) These deposits varied greatly in the environments of deposition (Pl. 7, figs. 1, 2). Some made under standing water are included. Many of the exposures have been sampled during surveys of road materials by the Wisconsin Geological Survey, but owing to the great differences of results in different deposits no general summation is possible. Clearly defined deltas are present both near Manitowoc (Fig. 8) and southeast of Green Bay.

Glacio-Lacustrine Deposits

Glacial-lake beach deposits are abundant in Door County. They range from coarse stony gravel to sandy deposits depending on the distance from rock cliffs which furnished the material for many of them. Reference may be made to Goldthwait's (1907) descriptions of localities along the Green Bay shore. North of Green Bay near Point Sable (SE ¼ NE¼ sec. 1, T. 25 N., R. 21 E.) a gravel pit in very sandy material formerly exposed many clam and snail shells. This material is about elevation 600, 19 feet above Green Bay, clearly the Nipissing beach. Sandy lake deposits are found in several locations. Many exhibit oscillation ripple marks. The clays and silts which were deposited in deeper water than the sands and gravels were formerly exploited as a source for making brick. Good examples of sediments with annual laminations or varves were noted (1) on the north side of Manitowoc where Highway 42 reaches the shore, and (2) a short distance north of Two Creeks (sec. 11, T. 21 N., R. 24 E.). In many localities the layers have been greatly distorted by ice action or by slump (Fig. 1; Pl. 2).

The only chemical analyses of lake clays available (Ries, 1908; Buckley, 1901) show that they are highly dolomitic where fresh. At Manitowoc these clays were formerly used in making Portland cement, but for some time the clay requirement has been derived by washing the underlying gray till. The clay or winter layers are dark red, and the summer, silty beds are yellowish gray or gray.

Land Forms Caused By Glacial Deposition

Marginal Deposits or End Moraines

The outstanding land form of the area which was due to glaciation is the Kettle Interlobate Moraine. The dominant feature of this is the

Plate 3.—OVERRIDDEN CARY MORAINE AND VARVED CLAYS

Figure 1.—Cary interlobate moraine overridden by Valders ice. The areas of gray bouldery Cary till surrounded by the red Valders till were long a puzzle. This portion of the Interlobate Moraine in SE ¼ SE ¼ sec. 19, T. 22 N., R. 24 E., here shown indicates little alteration by overriding the Valders ice. The Valders till was so thin that its identity has been destroyed by soil-making processes.

Figure 2.—Contact of Cary till and varved clays of Early Lake Chicago. The Cary till shown near the bottom of this exposure at the original Forest Bed locality in sec. 11, T. 21 N., R. 24 E., is higher in clay and silt than it is in most of the region. There is no evidence of any unconformity before the varved clays above were deposited in Early Lake Chicago.

Plate 4.—TWO CREEKS FOREST BED

Figure 1.—Upright stump of spruce tree in silt of Early Lake Chicago, Sec. 2, T. 21 N., R. 24 E. Above the stump is sand deposited in front of the Valders ice in the early phase of Later Lake Chicago (pro-Valders deposit). Top of the lake cliff is Valders till. The sub-Forest Bed varved clays of Early Lake Chicago are here largely concealed by slump.

Figure 2.—Clays of Early Lake Chicago intruded into later deposits. Same locality as Figure 1. The red Valders till has a nearly level base on top of sand, silt, and clay of pro-Valders age. The logs of wood in this photograph are all driftwood. At this place the underlying clays of Early Lake Chicago were so much distorted by the pressure of the Valders ice that small anticlines or domes were intruded into the lake sediments and truncated by the Valders ice.

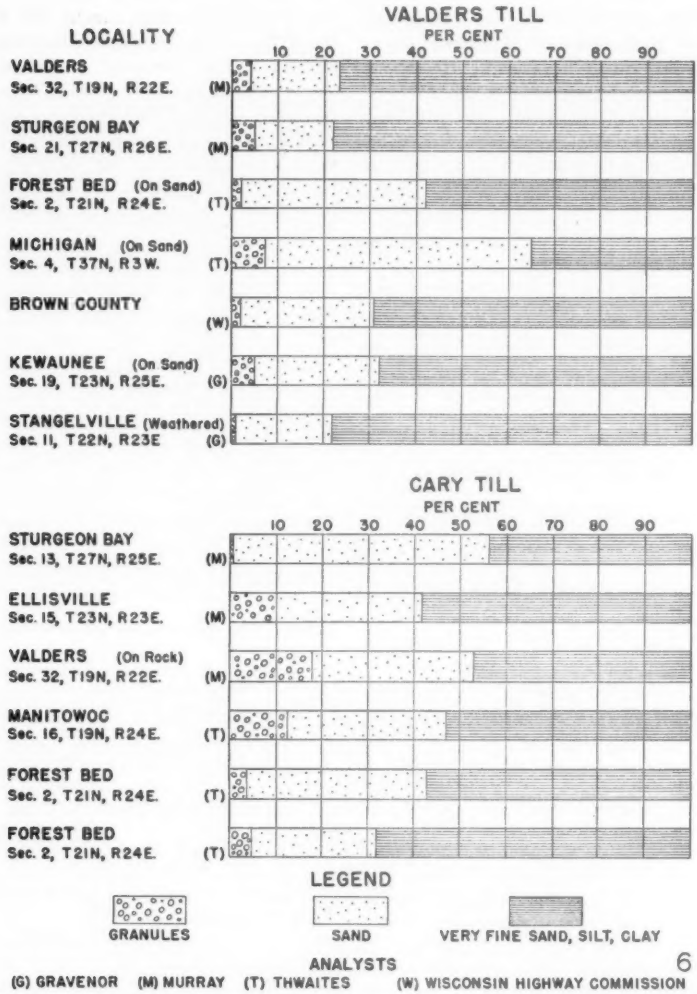

FIGURE 6.—MECHANICAL COMPOSITION OF VALDERS AND CARY TILLS OF NORTHEASTERN WISCONSIN AND PART OF SOUTHERN MICHIGAN

extreme irregularity of the surface with many enclosed depressions known as kettles. Relief exceeds 100 feet. In the southernmost part of the area, near St. Nazianz the red-till cover is absent, and the topography of the moraine is best exhibited. Farther north it is gradually obliterated by the overlying deposit, although there are locally conspicuous rough areas, for instance near Denmark (Pl. 3, fig. 1) and southeast of Whitelaw. Over considerable areas in eastern Brown County and north of Two Rivers the normal rough morainal topography is replaced by low knolls separated by shallow enclosed depressions or by irregular winding ravines. The relief of these minor knolls is in few places as much as 10 feet, but on the air photographs they are very prominent giving, especially in certain kinds of crops, a mottled

appearance (Pl. 1, fig. 2). Striking conical moulin kames are abundant in some areas. A good example may be seen from U. S. Highway 151 in sec. 16, T. 19 N., R. 21 E. As a whole, however, this region lacks the spectacular glacial topography which is displayed by the Kettle Moraine farther south.

Ground Moraine

Ground moraine may be classed as the areas of till which are devoid of distinct recognizable end moraines or drumlins. In the Door Peninsula the ground moraine may be divided into (1) gently rolling to flat areas of red glacial till, and (2) somewhat similar topography with very thin drift over the bedrock. Type (2) is found mainly north of Sturgeon Bay, although there are examples not far northeast of Green Bay along the top of the Niagara escarpment. The red till cannot be recognized in most of these areas of thin drift.

Drumlins

The areas of ground moraine are diversified in some localities by distinct drumlins. Kowalke (1952) described some of the drumlins of northern Door County. There are prominent drumlins just south of Sturgeon Bay, one of which exhibits gravel on its crest (Pl. 2, fig. 2). Farther south most drumlins were mapped from the air photographs. Many that are mantled by red till are very imperfect in form. Alden noted several drumlins near Valders and Chilton, and Thwaites first observed the prominent examples in southeastern Brown County in 1922. Most of these have little or no recognizable red till on top. Almost all the drumlins trend either nearly due south or to the east of south and have at least a core of gray till. Only west of Chilton in the area outside the red till do they have a southwesterly direction. Few are so conspicuous as the large drumlins either south or west of the area under discussion. By reason of their low relief few drumlins show in the topographic map with a contour interval of 50 feet (Pl. 8).

Outwash

Deposits of glacial streams are not so abundant as in areas to the south and west. Such deposits are divisible into (1) Cary or pre-red till and (2) Valders or post-red till. Those of (1) are not recognized in many exposures but are chiefly inferred from the topography combined with well records, many of them furnished by the State Board of Health. The cover is red glacial till. The western area of buried outwash south of Brillion is very clear cut in flat-topped topography between definite kettle holes, although the limits inferred are uncertain in places. The Brillion village well No. 1 shows 30 feet of sand and gravel beneath 15 feet of red till. Below this outwash is 70 feet of lake clay which extends to bedrock. In many places thin layers of sand and gravel within till are neglected since they are probably either local or possibly are transported masses. The area of overridden outwash west of Manitowoc is by no means as definite as the western area. In the vicinity of Silver Lake (sec. 34, T. 19 N., R. 23 E.), however, relations are clear, although no exposure of the buried outwash was discovered. Outwash younger than the red glacial till is distinguished on the soils maps as "Fox" and on older maps also as "Plainfield." The boundaries determined by the soils men were followed on the map. The deposits are mainly sand with minor amounts of gravel. Large commercial pits are, or once were, operated near Casco. Here lake clay underlies the outwash. The major area of outwash extends south from Casco almost to Two Rivers. There are also many smaller areas. Almost all the topography is typical pitted plain with many small kettles although there are no prominent lakes. Much of the outwash which post-dates the red till has been eroded into terraces.

Eskers

The most spectacular eskers of the region here described are (1) north of Brillion, (2) north of Casco extending into Door County, (3) in the vicinity of Mishicot, and (4) near Francis Creek. A ridge of gravel, nearly all now excavated, west of Maribel is also mapped as an esker although it is closely associated with the Interlobate Moraine. Some of these ridges attained an elevation of almost 50 feet above the surrounding country and are visible at considerable distances across the near-by glacial till plains. It is thought that all are older than

the red till, although a cover of that deposit is not everywhere in evidence. The Mishicot esker may extend south to Manitowoc beneath the red till cover and possibly joins some of the delta gravels of that locality (Fig. 8). The directions of the eskers check well with ice directions as shown by both striations and drumlins. The winding eskers must have been deposited when the ice sheet was stagnant. Allouez ridge, between Green Bay and DePere, is very straight and resembles the long sand eskers west of Fox River (Thwaites, 1943, p. 110). However, it is not here mapped because it was wholly submerged by the glacial lakes. Records of wells and test borings on the ridge show that there cannot be any continuous sand core. Sand was formerly excavated in several pits and shows in a railroad cut but of six test holes drilled in 1953 for water along the ridge only one discovered any sand. Most passed through 5–40 feet of red till resting on older lake clays. The following log of a recent water well shows the material near the center of the west flank of the ridge.

Partial log of Well No. 2, Wisconsin State Reformatory

Samples furnished by Layne-Northwest Company and examined by F. T. Thwaites

	Thickness, feet	*Depth, feet*
Clay, red, dolomitic (post-Valders)	5	5
Till, red, dolomitic (Valders)	5	10
Sand, fine, much silt, gray, dolomitic	10	20
Clay, silty, gray and red (varved?) dolomitic	75	95
Till, gray, dolomitic (Cary)	55	150
Bedrock		

The log of Allouez village well No. 2, given later in this report, shows a similar section. The discontinuity of the sand may be explained by overriding by glacial ice of the Valders substage.

Lake Deposits

Land forms caused by lake action are found in low lands adjacent to the present shore lines of Green Bay, Lake Michigan, and Lake Winnebago, as well as in low tracts along Manitowoc River. The studies of the several outlets of glacial lakes now described under the heading of Glacial History have greatly extended the areas known to have been submerged by glacial lakes which did not leave marked shoreline features. For this reason the present geological map does not agree with that of Thwaites' published in 1943. Some of what is there called Valders outwash covered with dunes south of Oconto Falls is probably shore deposits of Lake Oshkosh. Much of the former lake bottom along the shore of Lake Michigan is covered by sand dunes, which to a large extent conceal the bars and cliffs formed by the waves. Good examples of dunes occur on Rawley Point at Point Beach State Forest and near the Sturgeon Bay Ship Canal. Only the larger areas are shown on the map.

The present-day inland lakes of Door County appear to owe their origin to the obstruction of depressions in the rock surface by lake-beach deposits. For long distances the former lake limit is marked by a prominent cliff. Good examples occur at Robert La Salle county park near the southeast corner of Door County, as well as on Rawley Point. Goldthwait describes many farther north. Most of the shore line (Fig. 16) is at an elevation 17 to 20 feet above that of the present Lake Michigan and represents the Nipissing level. At the northern end of Door Peninsula Goldthwait measured beaches up to almost 100 feet above that datum (Lake Algonquin). Many of these in the more exposed localities open to the full force of the waves of the predecessors of the present lake are cut in the bedrock. Near Ephriam there are well-developed caves along the former water levels. On the Green Bay shore beach features are neither so common nor so spectacular. Locally bars of gravel were observed, as at Egg Harbor (Goldthwait, 1907, p. 89). Mapping of the beaches is based on air photographs and topographic maps supplemented by the observations of Goldthwait (1907), Shrock (1940), and Kowalke (1938, 1946).

History of Glaciation

Introduction

The history of glaciation in the Door Peninsula is largely inferred from the results of well drilling in localities where the glacial drift is

relatively thick. Since deposits of the several ice advances examined in shallow excavations are known to have distinct physical characteristics the stratigraphic approach can be used to interpret samples taken in water wells. The following succession of deposits is compiled from various localities.

General stratigraphic column of glacial and postglacial deposits of the Door Peninsula

Postglacial or Recent
- Dune sands
- Beach and lacustrine deposits, sand, gravel, gray clay of present lakes; marl and muck of enclosed waters
- Beach and deep-water sediments of Lake Nipissing
- Unconformity with very low level of Lake Michigan, 230 feet above present sea level
- Beach deposits and varved clays of Lake Algonquin (Later Lake Chicago, Toleston, and possibly Calumet levels)

Wisconsin glacial stage
- Valders substage
  - Till, red, very high in clay and silt; some sand and gravel
  - Sand, gravel, silt, and clay deposited in lakes in front of advancing Valders ice
  - Organic remains, trees, logs, mosses, etc. (Forest Bed)
  - Beach deposits and varved clays with some sand of Early Lake Chicago (Glenwood and Calumet levels)
- Cary substage
  - Till, gray, moderate clay content (except along Lake Michigan), including masses of varved clay, sand, and gravel in part plowed up from older deposits
  - Lake deposits, mainly varved clay, not known at surface
- Pre-Cary substage or stage
  - Till, brown or red, not known at surface

Bedrock (Paleozoic)

The following section on the north shore of Lake Winnebago shows a very common succession of drift deposits.

Partial Log of Sensenbrenner well (SW ¼ NE ¼ sec. 22, T. 20 N., R. 18 E.

Samples furnished by J. J. Faust and examined by F. T. Thwaites

	Thickness, feet	*Depth, feet*
Clay, red, dolomitic (Later Lake Oshkosh)	18	18
Till, red, dolomitic (Valders)	11	29
Clay, gray, dolomitic (Early Lake Oshkosh)	75	104
Till, gray, dolomitic (Cary)	20	124
Bedrock		

The following wells near Green Bay exhibit the most nearly complete recorded sections:

Partial Log of Well No. 1, Town of Preble, corner Henry and Deckner Avenues (Elevation 601. SW ¼ SE ¼ sec. 32, T. 24 N., R. 21 E.)

Samples furnished by Jos. Egerer and examined by F. T. Thwaites

	Thickness, feet	*Depth, feet*
No sample (probably Valders till)	10	10
Clay, red, dolomitic (pro-Valders)	10	20
No sample	10	30
Clay, gray, dolomitic (Early Lake Chicago)	50	80
Till, gray, dolomitic (Cary)	50	130
Clay, pink, dolomitic (pre-Cary)	5	135
Gravel, coarse (pre-Cary)	5	140
Till, brown, dolomitic (pre-Cary)	11	151
Bedrock		

Partial Log of Well No. 2, Town of Allouez. French claims corresponding to W. part sec. 12, T. 23 N., R. 20 E.

Samples furnished by Fasbender Brothers and examined by F. T. Thwaites. This well is about 2¾ miles southwest of No. 1, Preble, and 60 feet higher, on Allouez Ridge

	Thickness, feet	*Depth, feet*
Till, red, dolomitic (Valders)	10	10
Sand, fine, gray (core of ridge) (late Cary)	20	30
Till, gray, dolomitic (Cary)	35	65
Sand, fine, gray, much silt (Cary)	5	70
Till, sandy, gray, dolomitic (Cary)	15	85
Clay, sandy, pink, dolomitic (pre-Cary)	45	130
Till, much clay, pink, dolomitic (pre-Cary)	35	165
Bedrock		

Pre-Cary Glaciation

The two wells given above are the only ones in which there is any definite suggestion of a till older than the Cary or "gray" till. Absence of any extensive remnants of such deposits is probably ascribable to erosion by the Cary ice whose till directly overlies the bedrock in the vast majority of recorded exposures and wells. This till may be of Tazewell age. It is probably related to the southwest-trending striae (Quarry, Valders, Brillion as reported by Shrock, Morrison, Sturgeon Bay, and Denmark) and to the southwest-trending overridden drumlins east of Fond du Lac (Alden, 1932, p.

40–41). If this suggestion is correct, a Lake Michigan lobe is indicated which extended much farther west than during the Cary substage.

Pre-Cary Interval

An interval between the probable pre-Cary till and the main body of Cary till is suggested in the above well records by clay and gravel. Since the land in the Green Bay depression probably was covered by a glacial lake, weathering of the till should not be expected if it is of Tazewell age. The gravel may be deltaic.

Cary Glaciation

The Cary substage of the Wisconsin stage of glaciation appears to have brought the bulk of the glacial drift of the area and caused most of the depositional land forms. It was probably equally important in erosion, forming the deep basin of Lake Michigan and the straight sections of the Niagara escarpment. It may also have excavated the enclosed rock basin west of Brillion where there is no suggestion in the well records of earlier till. As shown in Figure 6 the main part of the gray Cary till is comparatively low in silt and clay, for it is composed of broken up Niagara dolomite.

Direction of Ice Movement

The Cary ice movements, in so far as they can be measured, do not indicate the marked division into two lobes which is manifest farther south and was long ago described by Chamberlin (1877) and Alden (1918). Throughout Door County the motion was slightly east of south which indicates that ice coming from the north overrode the entire peninsula. Age equivalence is proved by the similarity of trend of striae and drumlins. Farther south, in Manitowoc County, the movement appears to have been nearly due south. In the quarry at Valders (Fig. 7) the south-trending striae occur alone under the gray till which is preserved on both north and south sides of the hill. Farther west, along the highest part of the Niagara escarpment, the drumlins near Chilton exhibit a marked radiation, trending in part east of south and in part west of south. The diversion to the west is clearly the result of faster motion of the ice in the Lake Winnebago lowland. At Brillion the striae under the gray Cary till trend 140°, well to the east of south, although at most points on the escarpment, and in the lowland to the west, motion was west of south.

Formation of End Moraines

Moraines deposited at the border of the waning Cary glacier, whether due to a balance between forward motion and melting or to a readvance, are here termed end moraines. Northwest of Chilton two distinct moraines mark margins of the ice on the backslope of the escarpment and are approximately normal to the ice-flow direction. On the west, however, these merge into rough ice-margin topography, with abundant kames, which lies along the face of the escarpment which bounded ice that had melted back into the Green Bay-Winnebago lowland. The low smooth ridge forming the divide between Lake Winnebago and Fox River is apparently caused by the southeastern continuation of the Athelstane Moraine of the report of 1943. It can probably be correlated with the marked moraine just south of Morrison which trends normal to local ice flow. A southeasterly ice flow is demonstrated here by an extensive field of drumlins. The wide moraine from Shirley north to Door County is younger and shows that the ice melted back into the lowland now occupied by Green Bay.

Sand Ridges of Green Bay Lowland

When the ice was decaying after the formation of the Morrison Moraine four distinct ridges of sand, gravel, and silt were formed in crevasses which paralleled the southwesterly ice flow when the lobe was active. Allouez ridge between Green Bay and De Pere and a small hill in the eastern part of Green Bay are the only ones in the area (Pl. 8). Exposures of the material of these ridges are for the most part rather poor, but it is clear from the disturbance of the upper part of the stratified deposits that they are of Cary age.

Formation of the Kettle Interlobate Moraine

With the exception of the Brillion esker, adjacent drumlins, and two occurrences of striae east of Denmark, there is little suggestion of a marked division into Green Bay and Lake

Michigan lobes; nevertheless, the Interlobate Moraine can be traced throughout most of the district here mapped. As the ice margin was melted back toward the north a re-entrant angle developed. The moraine near St. Nazianz is very complex both in topography and internal composition. The complexity of deposits

The extreme variability of materials shown by these logs is doubtless a combination of frequent changes in position of the ice margin and of plowing up of older stratified deposits. The very prominent spur of the moraine which extends north and northwest toward the pitted outwash south of Brillion was at one time be-

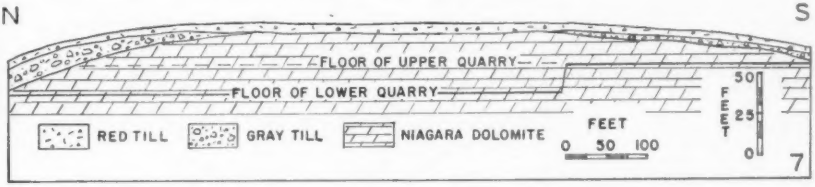

FIGURE 7.—SKETCH OF FACE OF QUARRY AT VALDERS, WISCONSIN
Showing relation of drifts to bedrock

within the Kettle Moraine is shown in the following logs of two wells drilled on the side of a kettle hole near the south boundary of the map.

Partial Logs of wells at Salvatorian Seminary, St. Nazianz, Wisconsin

Samples furnished by Layne-Northwest Company and examined by F. T. Thwaites

Well No. 1

	Thickness, feet	*Depth, feet*
Clay, sandy (no sample)	5	5
Sand, clean, gray (no sample)	6	11
Clay and sand (no sample, probably till)	34	45
Sand, very fine, gray, dolomitic	20	65
Gravel, coarse (may include till to 70)	15	80
Till, gray, dolomitic	15	95
Gravel, coarse (water)	8	103
Till, gray, dolomitic	22	125
Gravel, fine	5	130
Till, gray, dolomitic	40	170
Gravel, coarse (water)	6	176
Bedrock		

Well No. 2

	Thickness, feet	*Depth, feet*
Sand, fine, light gray	55	55
Till, gray, dolomitic	40	95
Sand, coarse to medium, light gray	10	105
Silt, sandy, light gray, dolomitic	10	115
Sand, very coarse to medium, light gray	5	120
Till, gray, dolomitic	5	125
Gravel, stony	15	140
Sand, fine to small pebbles	30	170
Gravel, some sand	5	175
Bedrock		

lieved by Thwaites (Alden, 1932, p. 44) to be the main continuation of the Interlobate Moraine, but is here interpreted as a re-entrant-angle deposit between ice in the low ground now occupied by Manitowoc River near Quarry and the swampy lowland east of Chilton. If this is the case, the buried outwash is younger and may be the product of drainage into a lake of Cary age near Quarry, west of the Interlobate. It is clear that the Interlobate Moraine proper swings northeast near St. Nazianz as mapped by Alden (1918). Its northward continuation is less definite. The morainal belt which passes through Whitelaw to Denmark terminates abruptly near Luxemburg. The marked parallel ridge through Bolt joins this moraine at both ends.

Lake-Shore Moraines

The morainic belt which extends from Sturgeon Bay south along the lake shore to Two Rivers could be interpreted either as (1) a phase of the Interlobate Moraine or (2) a retreatal moraine or moraines of the Lake Michigan Lobe. The eastern part may extend out into Lake Michigan at Two Rivers, but the western portion is probably of the same age as the two small moraines along the lake shore south of Manitowoc. The topographic expression of all these moraines is weak over large areas because of a heavy cover of red Valders till. Sand is found below Valders till in southeastern Door County.

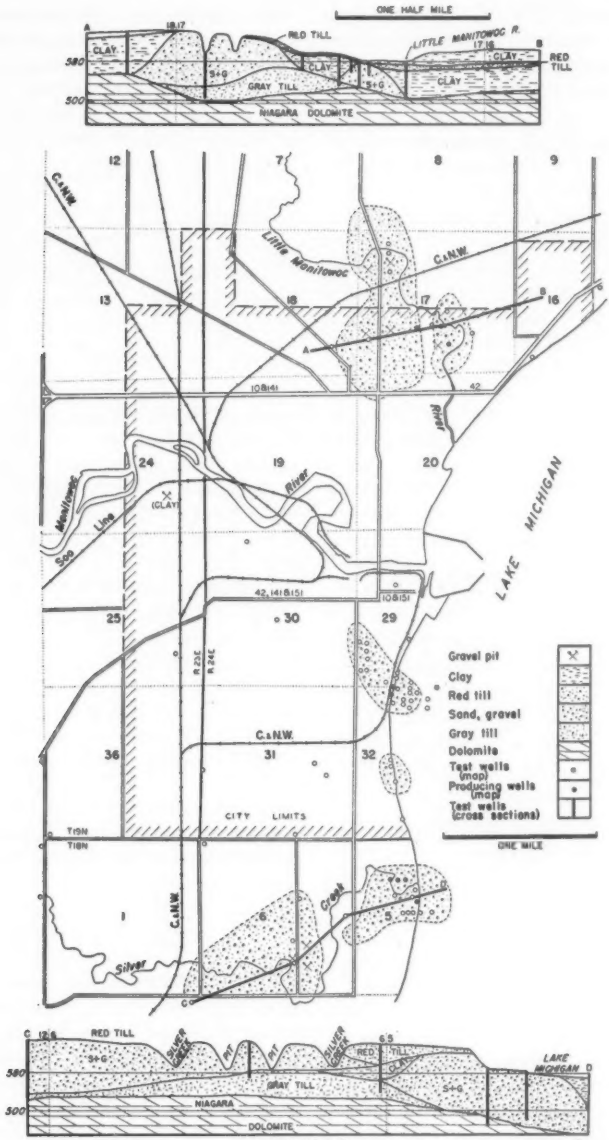

FIGURE 8.—MAP AND CROSS SECTIONS SHOWING CARY GRAVEL DEPOSITS INTERPRETED AS DELTAS DEPOSITED IN EARLY LAKE CHICAGO
Data furnished by W. G. Kirchoffer, Engineer, and by drilling contractors

The area of delta moraine at Manitowoc (Fig. 8) was known from both gravel pits and test borings for water. Delta origin in front of the retreating ice was deduced from the prevailing fore-set bedding which dips west. Lack of agreement between adjacent test borings is also indicative of inclined bedding. Although these deposits are located along the line of the Mishicot esker they are probably not related in origin. The lake level seems to have been about 640 feet elevation, but overriding by the later Valders ice plus deposition both pre-Valders and post-Valders of lake sediments have obscured relations. In parts of the area no red Valders till could be found above the delta gravel, possibly owing to erosion by a post-Valders glacial lake. Ice contortion of the upper layers of the deltas was observed in several of the pits which are now abandoned and filled.

Moraines of Door County North of Sturgeon Bay

North of Sturgeon Bay many small patches of morainic topography have been mapped, in part from air photographs but largely from the soils map which designates them as "Miami gravelly loam". A peculiar feature of these tracts is that many of them, despite their typical knob and kettle topography, trend east of south or parallel to the direction of the striae. In fact a few look like true eskers, but the breadth and irregular outline do not support this interpretation. Some of these areas are true moraines transverse to ice movement. Good exposures of the material are rare.

Soils of the Cary Drift

Where, for instance in southern Calumet County, the Cary drift forms the surface, the soils are much more stony than they are in areas of the red Valders till. Within the area covered by the Valders glaciation there are considerable areas where the later till was so thin that soil-making processes have destroyed its identity. Within most of these areas there are small isolated patches of the younger red till. It did not prove practicable to map many areas where the drift is very thin. The largest area of thin drift is north of Sturgeon Bay in Door County. It is the location of many large cherry and apple orchards, but probably climatic control is more important than soil. The adjacent bodies of water delay the spring and hence prevent late frosts. On the maps of Door and Kewaunee counties (Whitson, 1914; 1919) the areas where the Valders red till cannot be found are called "Miami". On later maps (Whitson, 1925; 1926; 1929) this name was replaced by "Bellefontaine". Leaching of carbonates may extend to a depth of nearly 4 feet (Kellogg, 1930; Whitson, 1927). It is less in most of the area.

De Pere-Greenleaf Terrace

The Niagara escarpment from just east of De Pere to the vicinity of Greenleaf is fronted by a terrace of sand and gravel covered with red till (Figs. 2, 10). Summits of this much-eroded terrace reach 825 feet elevation, much higher than any part of the sand ridges described above. The relation of the deposits to the sand ridges to the northwest is not clear, but the gravelly material and the elevation of the highest parts suggest that deposition took place between the rock cliffs and the front of the waning Cary glacier. The outlet was to Early Lake Oshkosh in the Lake Winnebago basin which in turn drained to Wisconsin River near Portage at elevation slightly above 800 feet (Thwaites, 1943, p. 127–135). Several exposures in this belt display cross-bedding which dips south to southwest. In SE ¼ SW ¼ sec. 31, T. 23 N., R. 21 E., a bed of dolomite talus is contemporaneous with deposition of the sand and gravel with which it is interbedded. Even if the divide between the Fox River and the Manitowoc drainage, northeast of Sherwood, has been raised by deposition of the later Valders till, flow via this outlet to the east might have been blocked by ice which lingered in the Lake Michigan basin.

Early Lake Oshkosh

A glacial lake must have formed in the Fox River valley in front of the retreating Cary ice just as one did in front of the later Valders ice. Evidence of such a lake has been given by Thwaites (1943, p. 132–135). In the area here

FIGURE 9.—CONTOUR MAP OF ERODED TERRACE NEAR GREEN BAY

described many well logs, some given above, disclose lake clays between the Valders and Cary tills. The flatness of much of the till plain throughout the lower Fox River valley and scattered exposures demonstrate the same thing. The only part of this overridden lake plain that can be shown on Plate 8 is that which is higher than the area submerged by the later lake. The history of the drainage of Early Lake Oshkosh across the Niagara Escarpment is unknown. It may have paralleled that of the later lake, and when the low cross-valley at Sturgeon Bay was uncovered this lake merged with the waters enclosed in the Lake Michigan basin.

Early Lake Chicago

The ponding of water in front of the waning Cary ice in the basin of Lake Michigan has long

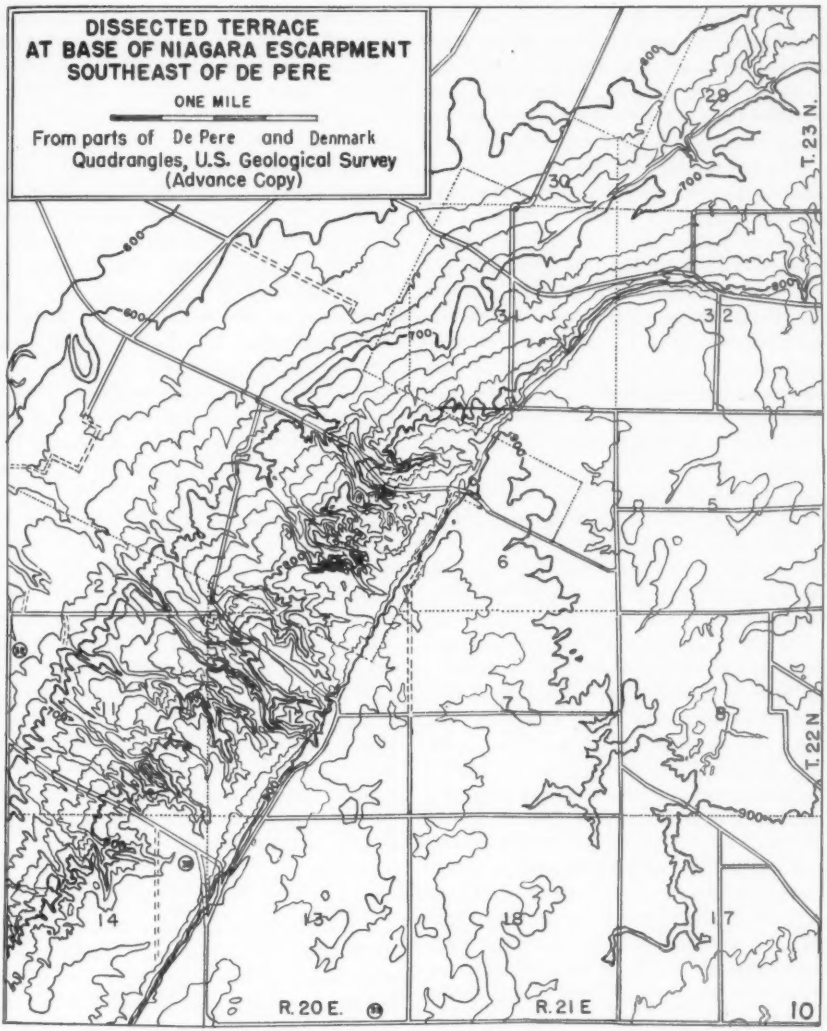

FIGURE 10.—CONTOUR MAP OF ERODED TERRACE NEAR DE PERE

been known. Clays which occupy the same stratigraphic position as those of Early Lake Oshkosh are abundant throughout the district. (Figs 8, 12, 13). These clays, most of which are varved, are widely exposed along the shores of Lake Michigan, for instance at the Forest Bed, the clay pit in Manitowoc (Pl. 2, fig. 1), and at many other localities. The studies of Bretz (1951) and Hough (1953a; 1955) indicate that the outlet to Desplaines River west of Chicago had been eroded down to the Calumet or 40-foot level prior to draining of the lake by the freeing of the Mackinac Straits region from glacial ice. How low the water level then fell in the interval between Cary and Valders glaciations is unknown.

Two Creeks Interval

Deposits of clay, silt, sand, and gravel formed during the Two Creeks interval between

the Cary and Valders substages of the Wisconsin glaciation have been identified in many localities throughout the region. It is difficult in many places, however, to determine what portion is late Cary age and to what extent the deposits were laid down during the advance of the Valders glacier. In the vicinity of Two Creeks imperfectly varved red and gray glacial-lake clay more than 10 feet in thickness lies directly on a till regarded as Cary (Pl. 3, fig. 2; Pl. 5, fig. 2). Locally this clay is overlain by a few feet of sand and silt (Fig. 12). On top of the clay and the sand there grew a considerable forest of spruce trees; the stumps of many trees still remain in the position of growth. A poor exposure of the Forest Bed was found in sec. 16, T. 24 N., R. 25 E., and logs were seen in the red till and sand northeast of Green Bay (SW ¼, SW ¼, sec. 23, T.24 N., R.21 E.). Logs of wood have also been reported in wells near Forest Junction (Lawson, 1902), near Bellevue, and elsewhere in the region of thicker drift. The paleobotany of the Forest Bed was first studied by Wilson (1932; 1936). In 1953 W. L. Culberson (1955) made extensive collections.

The wood consists of one or possibly two species of spruce[3] of which *Picea mariana* is the more abundant. Some traces of balsam (*Abies balsamea*) were found above the main part of the Forest Bed (Wilson, 1936, p. 320). Growth rings, by becoming much thinner in the last 20 years of life of the trees, demonstrate that the climate deteriorated. One log was an exception; it may have been transported by ice from another locality. Ring successions with an observed maximum of 142 differ, which also suggests that certain logs came from upland trees that grew on a different soil where moist years increased rate of growth instead of reducing it as was the case of the dense wet soils of the original Forest Bed locality. Immediately below the peat and moss of the Forest Bed, the clay contains mollusks which were identified by F.C. Baker (Wilson, 1932, p. 39), who concluded that the habitat of these was a very moist soil. Seven species were identified.

Three species of mollusks were found (Wilson, 1932, p. 39–41) near the top of the varved clay below the original locality in sec. 11, T. 21 N., R. 24 E. These were determined by F. C. Baker as *Fossaria dalli* (Baker), *Pupilla muscorum* (Linnaeus), and *Succinea avara* (Say). The forms lived in wet mud and forest. Higher up land forms appear including *Vertigo ventricosa* (Morse), *Bryum cyclophyllum* (Schwaegr.) and still higher up *Gyraulus circumstriatus* (Tyron), *Fossaria parva* (Lea), and *Pisidium* sp. The last three are aquatic and indicate the submerging of the deposit before the ice reached that region.

Wilson (1932, p. 38) had the mosses identified by L. S. Cheny who distinguished eight species of mosses whose present range is north of this locality. Wilson concluded that the splintering of the ends of the logs indicated "a violent twisting and bending of live trees before they were felled." Comment on this is given later. Culberson (1955) also found eight species of mosses with the aid of W.C. Steere: *Brachythecium salebrosum*, *Calliergon turgescens*, *Campylium stellatum*, *Ditrichum flexicaule*, *Drepanocadus uninatus*, *Abietinella abietina*, and *Tortella tortuosa*. All are "associated with floras of more northern affinities".

Pro-Valders Lake Deposits

Wilson (1932, p. 41) first called attention to a thin layer of sediment above the Forest Bed which is older than the overlying red till. In his later work he discovered more of this deposit, for he remarks (1936, p. 322): "There is a reappearance of the *Difflugia* shells, indicating the presence of shallow standing water. Then there was a general influx of water over the area and the forest was drowned and buried several feet deep in lacustrine sediments." Unfortunately the level of Lake Michigan was very low (Fig. 11) at the time of Wilson's later work. Much better exposures owing to a high lake level showed the present writers that this sediment is thicker than Wilson thought. Plates

[3] In 1956 Mr. Peter L. Thiel, a student at the University of Wisconsin, submitted a cross section from a log 11½ inches in diameter. This log was reported to have been found in an upright position 20 feet below the surface in brown varved clay at building 79 of the Kimberly Mill of Kimberly Clark Corp. Through the courtesy of Prof. R. W. Lounsbury, Purdue University, the wood was determined in thin section and found to be elm, probably red elm. The kind of wood, which has never before been reported from the Forest Bed, plus the unusual fresh and sound character of the specimen throw great doubt on its age. It might have been buried by previous building operations.

4 and 5 and Figure 12 show these pro-Valders lake sediments, which locally exceed 10 feet in thickness at this locality and contain much drift-wood. Owing to ice disturbance thickness is very variable. A lens, several feet thick and also highly distorted, is exposed at the southern locality east of Two Creeks. Work by the present authors disclosed the presence of extensive deposits of gravel, sand, silt, and red clay which in many localities display oscillation ripple marks. The material is all highly dolomitic, which indicates derivation from fresh drift that contained considerable red iron oxide (Pl. 7, fig. 1). At first it was thought that the absence of any remains of the Forest Bed between these deposits and the Valders till above was due to glacial erosion and that the irregular contact shown in Plate 6 near Algoma and in exposures near Two Rivers indicated an unconformity. Later study of the northern Forest Bed locality, however, showed this view is incorrect and that these lake deposits lie above the horizon of the Forest Bed. Besides these exposures good examples were found in the vicinity of Kewaunee; most of these are in or near valleys of streams which enter either Green Bay or Lake Michigan. The following well log indicates the succession.

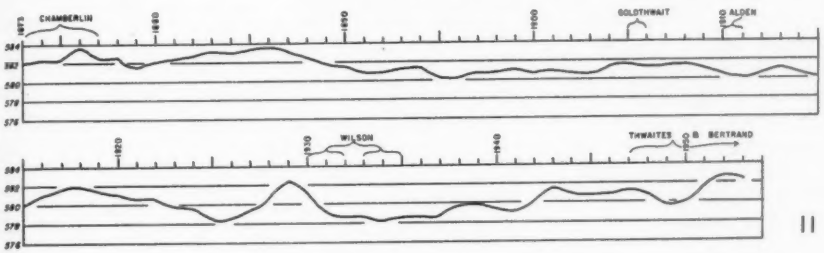

FIGURE 11.—SUMMER MAXIMUM LEVELS OF LAKE MICHIGAN FROM 1873 TO 1953
Showing time of different surveys. After U. S. Lake Survey. Chamberlin, Goldthwait, and the present authors were favored by high levels which caused much shore erosion.

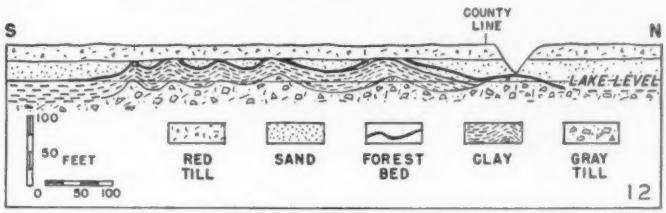

FIGURE 12.—EXPOSURES OF THE TWO CREEKS FOREST BED AND ASSOCIATED DEPOSITS
In shore cliff of Lake Michigan near the line between Kewaunee and Manitowoc counties. The folding of these beds can be ascribed to glacial shove. *See* Plates 4 and 5.

Partial log of Liebman Packing Company well, Green Bay, Wisconsin. (on ridge south of Highways 54 and 57 in sec. 33, T. 24 N., R. 21 E. Elevation 660)

Samples furnished by Jos. Reynen and Jos. Egerer and examined by F. T. Thwaites

	Thickness, feet	*Depth, feet*
Sand, fine, much silt, yellow gray, dolomitic	30	30
Clay, silty, gray, dolomitic	100	130
Clay, silty, red, dolomitic	10	140
Clay, gray, and pink gray, dolomitic (varved?)	90	230
Till, gray, dolomitic (Cary)	5	235
Bedrock		

At this locality the Valders till has been removed either by erosion or by grading. The horizon of the Forest Bed is indeterminate but may be at depth 140 feet. This would make the remainder of the clay below a deposit of Early Lake Chicago.

Pro-Valders deposits were explored for water by 21 test holes drilled under direction of the

McMahon Engineering Co. Of these, 9 are located in the Y of Highways 54 and 57 just west of the Brown County Farm in SE ¼ SW ¼ sec. 27, T. 24 N., R. 21 E. Only one of the tests, at this place which range from 30 to 80 feet in depth, reached the Cary gray till. They passed through clay, silt, sandy silt, and sand with only a small amount of gravel. Holes a short distance apart showed entirely different successions of materials. Those given below were the only ones in which Valders red till was found. Another hole half a mile north and half a mile east passed through 34 feet of red Valders till into coarse gravel and sand to a depth of 125 feet where it reached the Richmond shale. During 1955, 12 test holes were drilled by Layne-Northwest Company for the Town of Preble, under direction of McMahon Engineering Company. These are farther south and ranged from 20 to 135 feet in depth. Only a few found any delta gravel or sand of pro-Valders age. Most of them seem to have been located too low for these deposits since they found lake clay and gray Cary till. A few lenses of gravel and sand within or just above the till were encountered. Most of these contained much silt and no large amount of water. There has been too much postglacial erosion of this portion of the deposit to allow the presence of much water.

Log of Test Hole No. 8, Town of Preble

Samples furnished by Jos. Reynen and examined by F. T. Thwaites

	Thickness, feet	*Depth, feet*
Sand, very fine to coarse, silty, gray, dolomitic	5	5
Till, pale red, dolomitic	20	25
Till, pink-gray, stony, dolomitic	9	34
Sand, fine to very coarse, gray, dolomitic	6	40
Gravel, fine, stony, gray, water	15	55
Gravel, like above with some clay, red, dolomitic	2	57
Sand, very fine to fine, silty, gray, dolomitic	17	74

Log of Test Hole No. 9, 10 feet north of No. 8

Samples furnished by Layne-Northwest Co. and examined by F. T. Thwaites

	Thickness, feet	*Depth, feet*
Sand, fine, light brown	5	5
Till, gray pink, dolomitic	15	20
Gravel, fine, much clay, pale red	15	35
Sand, very coarse to fine, light gray, dolomitic	5	40
Gravel, coarse, stony, some sand layers, water	15	55
Sand, fine, light gray, dolomitic	20	75
Clay, pale red, dolomitic	5	80

The extreme irregularity of these deposits is ascribed in part to the delta fore-set bedding

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PLATE 5.—TWO CREEKS FOREST BED

FIGURE 1.—Valders till resting on pre-Valders disturbed sediments. The base of the red Valders till, which contains some lenses of stratified material, is here almost level. At the right is a lens of pro-Valders sand containing driftwood spruce logs. The sediments to the left, where the bedding is vertical, are of pro-Valders age. In the middle is a mass of gravel of unknown source. In many places ice disturbance is so extensive that the original bedding cannot be distinguished. Sec. 2, T. 21 N., R. 24 E., south of photographs of Plate 4.

FIGURE 2.—Terminus of sand lens with included driftwood logs. The material above the sand is red Valders till. Below the sand are the varved clays of Early Lake Chicago. At the far right these rest on Cary gray till. Effect of ice shove is evident.

PLATE 6.—CONTACTS OF VALDERS TILL WITH PRO-VALDERS LAKE SEDIMENTS

FIGURE 1.—Valders till filling hollow in very fine silty sand of pro-Valders age. Such hollows were at first taken for erosion valleys, but there is no evidence of the Forest Bed at this horizon. It is much more probable that they are due to ice shove. NE ¼ NE ¼ sec. 27, T. 25N., R. 25 E., Algoma, Wisconsin.

FIGURE 2.—Valders till filling depression in surface of pro-Valders lake deposits. The filled depression is larger than that of Figure 1 and may be a kettle caused by melting of an iceberg stranded in the bottom of Later Lake Oskosh. There is no evidence of ice disturbance. SE ¼ NE ¼ sec. 4, T. 23 N., R. 21 E., east of Green Bay, Wisconsin.

FIGURE 2

FIGURE 1

TWO CREEKS FOREST BED

FIGURE 1

FIGURE 2

CONTACTS OF VALDERS TILL WITH PRO-VALDERS LAKE SEDIMENTS

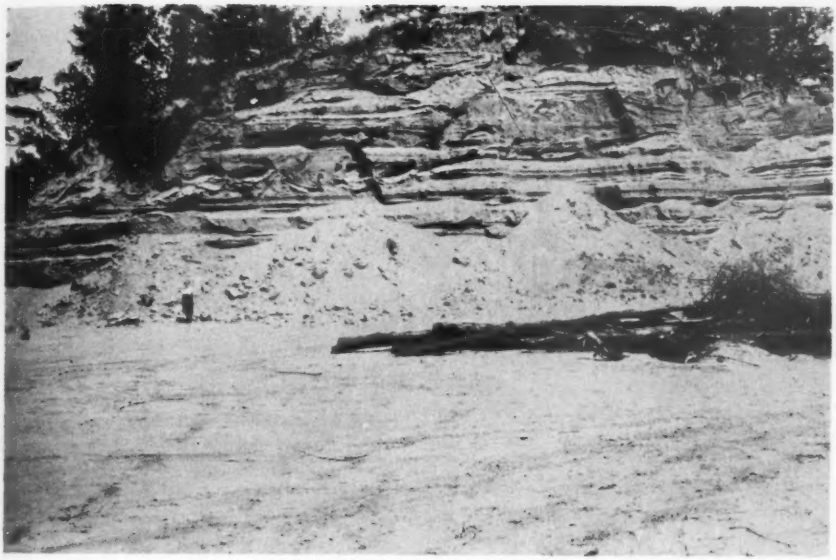

FIGURE 1

FIGURE 2

PRO-VALDERS DEPOSITS NEAR GREEN BAY

which dips southwest and in part to disturbance by Valders ice. Large gravel pits in these deposits have been operated in sections 33 and 35, T. 24 N., R. 21 E., and in sections 2, 3, and 4, T. 23 N., R. 21 E. Their topography is shown in Figure 9.

History of Two Creeks Interval

The history of the interval between Cary and Valders glaciation includes three distinct phases. (1) When the ice retreated, varved clays were deposited in Early Lake Chicago (Lake Milwaukee of Martin, 1932, p. 454) which judging from the deltas and varved clay at Manitowoc then stood at the Glenwood stage about 60 feet higher than the present lake level or elevation about 640. During the retreat of the Cary glacier the drainage area of this lake was increased when the opening of Grand River across lower Michigan admitted water from lakes Maumee, Arkona, and Whittlesey (Bretz, 1951, p. 425). This increase in discharge lowered the outlet to the Calumet Beach about 620 feet. Although no definite trace of the early Calumet-level beach has been discovered in this area, the southern terrace of Manitowoc River at Manitowoc Rapids is about 36–40 feet above the present lake. This checks with the Calumet-lake level of late Valders time. (2) After the early Calumet level the waters were lowered to at least the present level, for otherwise the Forest Bed could not have grown as low as it is found. This low stage was named "Bowmanville" by Baker (1920, p. 69–77; 1926) but could better be termed Two Creeks (Fig. 13). It involved retreat of the ice front enough to clear not only the Straits of Mackinac with the lowland south of that, east of Little Traverse Bay, but also the entire St. Lawrence valley. Flow may have been via the drift-filled St. Davids gorge at Niagara Falls. If the land of that region lay as low as it did later, the level of Lake Michigan could have been several hundred feet below that of the present. During this interval the area was invaded by a spruce forest with associated mollusks, mosses, etc. However, there is no distinct soil profile beneath this plant growth. This is explicable by soil-drainage conditions which did not permit leaching in the impermeable lake clays and clay till, and by the comparatively short duration of the low-water stage compared with post-Valders time. (3) When the ice readvanced, during the Valders substage, the water level must have risen again to the Calumet outlet. During this rise of water level the trees seen in the exposures at Two Creeks were killed by submergence and burial by sand. Pro-Valders lake deposits were formed in both the Michigan and Green Bay lowlands. Depressions in the top of the sand and gravel of the Green Bay lowland, filled with red clay, silt, and red till, were observed in several localities and interpreted as filled kettle holes caused by buried ice blocks (Pl. 6, fig. 2). The control of top elevation of these deposits by a lake which stood at elevation not over 750 feet indicates a drainage outlet much lower than that over the Cary drift near Portage. Before deposition of the Valders till the lowland northeast of Lake Winnebago at the head of Manitowoc River was probably considerably lower than now and might have been the outlet.

Age of the Forest Bed

Until the discovery of the radiocarbon or Carbon 14 method of age determination, the

PLATE 7.—PRO-VALDERS DEPOSITS NEAR GREEN BAY

FIGURE 1.—Pro-Valders sands beneath Valders till. Approximately 35 feet of very fine sand, much of it with large oscillation ripple marks, overlain by red Valders till. There is no evidence of a Forest Bed. Ice disturbance is confined to shear planes with associated minor folding. SE ¼ SE ¼ Sec. 33, T. 24 N., R. 21 E.

FIGURE 2.—Topography on pro-Valders lake sediments. The locality is a short distance west of Figure 1, and both views are looking north. The extremely complex gulleying in the soft fine sands was mistaken by earlier geologists for a terminal or end moraine. However, there are no enclosed depressions, and the Valders till is confined to the highest points. *See* Figure 9 for contour map of this locality. It is believed that most of the erosion occurred in the interval between Algonquin and Nipissing stages of Lake Michigan since the larger valleys are cut across and filled by the Nipissing beaches.

age of the Forest Bed was only guessed at. A good detailed description of this method is that by Kulp (1952). In evaluating the results it is necessary to realize that the radioactivity of Carbon 14 in a sample of reasonable size is far below the background of stray radiation and cosmic rays. Elaborate precautions are needed

determinations are almost undoubtedly driftwood which may have been carried for some distance by water or ice. Wilson (1932, p. 37) recorded trees from a drier environment. Within a short period after the advancing Valders ice blocked the outlet or outlets at or near the Straits of Mackinac, water must have risen in

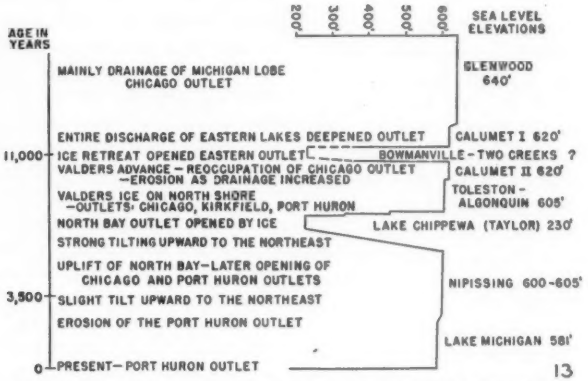
200'
300'
400'
500'
600'
SEA LEVEL ELEVATIONS
AGE IN YEARS
MAINLY DRAINAGE OF MICHIGAN LOBE CHICAGO OUTLET
GLENWOOD 640'
ENTIRE DISCHARGE OF EASTERN LAKES DEEPENED OUTLET
CALUMET I 620'
11,000
ICE RETREAT OPENED EASTERN OUTLET
BOWMANVILLE - TWO CREEKS ?
VALDERS ADVANCE - REOCCUPATION OF CHICAGO OUTLET -EROSION AS DRAINAGE INCREASED
CALUMET II 620'
TOLESTON - ALGONQUIN 605'
VALDERS ICE ON NORTH SHORE -OUTLETS: CHICAGO, KIRKFIELD, PORT HURON
NORTH BAY OUTLET OPENED BY ICE
LAKE CHIPPEWA (TAYLOR) 230'
STRONG TILTING UPWARD TO THE NORTHEAST
UPLIFT OF NORTH BAY—LATER OPENING OF CHICAGO AND PORT HURON OUTLETS
NIPISSING 600-605'
3,500
SLIGHT TILT UPWARD TO THE NORTHEAST
EROSION OF THE PORT HURON OUTLET
LAKE MICHIGAN 581'
0
PRESENT— PORT HURON OUTLET
13

FIGURE 13.—CHANGES OF WATER LEVELS IN LAKE MICHIGAN BASIN
After Hough

to reduce this background, but these are beyond the scope of the present paper. The probable error of all determinations is large: The recorded age determinations of wood and peat from the original exposure of the Forest Bed agree only fairly well (Table 1) (Arnold and Libby, 1949; 1950; 1951; Blau, 1953; Flint, 1956; Horberg, 1955).

To evaluate these data it is necessary to consider both instrumental errors and the fact that all the material may not be of the same age. The original five determinations were all of vegetation which grew at or near the point of collection. Doubtless its period of growth covered a considerable time, for Wilson (1936, p. 325) found tree trunks 142 years old. Although this alone could account for some of the scatter in results, which for these original measurements amounts to 1291 years, certainly instrumental errors, possibly including contamination with modern carbon either in the ground or in handling, were important. Much of the material since determined was not in position of growth. For instance all the Yale

the Michigan basin at least to the Calumet level or 40 feet above modern Lake Michigan. In the Fox River Valley the rise would have been the same until after the Sturgeon Bay depression was ice-covered and then would have followed in reverse the history of the lowering of Lake Oshkosh. This rise would flood the Two Creeks area. The spruce trees would then be killed by flooding. They must have been dead, although wet, when the ice completed their destruction.

Although there is no proof of vegetation growing on the shores of Early Lake Chicago before the drainage which allowed forests to grow on the lower slopes of clays and sands, there is no reason to doubt that trees continued to grow on the banks of later Lake Chicago which formed in front of the advancing Valders ice. Wilson reports pollen of upland trees (1932, p. 40; 1936, p. 130–131). If we neglect as an error the single determination of only 6400 years of specimen C419, the total scatter of all specimens is 2838 years, several times

TABLE 1—RADIOCARBON (C_{14}) DETERMINATIONS FROM FOREST BED

Lab No.	Collector	Remarks	Age in years	Reference
C308	Wilson	SE ¼ sec. 11, T. 21 N., R. 24 E.	10877 ± 740	Science, 1951, v. 113, p. 117-118
C365	Bretz	Same, root	11437 ± 770	Radiocarbon dating 1952, p. 88
C366	Bretz	Same, peat	11097 ± 600	
C536	Bretz, Horberg	Same	12168 ± 1500	
C537	Bretz, Horberg	Same, peat	11442 ± 640	
			Av. 11404 ± 350 (Scatter 1291 years)	
C800	Read	SE ¼ sec. 28, T. 21 N., R. 17 E. in red till	11471 ± 500	Science, 1954, v. 119, p. 139
			10241 ± 650	
			Av. 10856 ± 410 (Scatter 1230 years)	
C630	Lundsted, Read	Stump, Kimberly-Clark Mill	10676 ± 750	Science, 1952, v. 116, p. 678
Y141	Thwaites, Bertrand	Driftwood in red till	9930 ± 410	Science, 1953, v. 118, p. 1
		NE ¼ sec. 2, T. 21 N., R. 24 E.	11130 ± 350	Personal communication
			Av. 10530 (Scatter 1200 years)	
Y147	Thwaites, Bertrand	Driftwood in red till and sand SW ¼ sec. 23, T. 24 N., R. 21 E.	9330 ± 330	Science, 1953, v. 118, p. 1
			11940 ± 390 (Scatter 2610 years)	Personal communication
Y237	McMahon Eng. Co.	Menasha sewer trench	11690 ± 370	Personal communication
M	McMahon Eng. Co.	Kimberly sewer trench	11200 ± 600 (Center of log)	Personal communication
M	McMahon Eng. Co.	Kimberly sewer trench	11400 ± 600 (Rim of log)	Personal communication
C419	Thwaites, 1925*	Red till SE ¼ SE ¼ sec. 3, T. 20 N., R. 17 E.	5938 ± 300	Radiocarbon dating, 1950, p. 12; 1952, p. 88
			6864 ± 300	
			Av. 6401 ± 230 (Scatter 926 years)	
Mi-342	Zumberge and Spurr	Two Creeks	10700 ± 600 (Rim of log)	Science, 1956, v. 124, p. 669
Mi-343			10400 ± 600 (Center of log)	

Laboratories C = Chicago, Y = Yale, M = Magnolia Petroleum Co., Mi = Michigan

Total scatter 5767 years; neglecting C419, 2838 years. The scatter is too great for an accurate average of all observations.

* Original field notes by Thwaites of April 26, 1925, say: "On east bank (of Little Lake Butte des Morts) is about 15 feet of red clay. Top is laminated and apparently lake clay. Below appears like till. Logs up to 12" diameter slightly flattened, associated with green clay, peat, branches, etc., not a clear demonstration. Logs 15 feet below surface lie east-west." No excavation was made.

the variation in results for the original five specimens. This difference does not seem unreasonable for the time necessary for the ice to advance from the Straits region to northeastern Wisconsin, but may be wholly due to errors of the radiocarbon method.

The tree trunks found between Kimberly and Menasha are unique in showing flattening, a

phenomenon not recorded at Two Creeks or Green Bay.[4] Why this flattening is found only in the Menasha-Kimberly region is not known. The locality is below the level of glacial Lake Oshkosh, but trees could have grown there during the maximum of the Two Creeks recession.

Checks on Age Determination

A method of research relating age in years, depth of leaching, and rate of solution of soil carbonates was applied by G. H. Robinson in an unpublished thesis on the soils of the red-till region. Robinson used the direct or lysimeter measurement of the present-day rate of leaching. This he puts at 275 pounds of calcium-carbonate equivalent per acre per year. Using the known average depth of leaching and carbonate content of the till, and assuming that the bulk density of the soil is the same before and after leaching, he arrived at the age of about 9400 years for the soils on the Valders till. He used a carbonate content of 27.38 per cent, depth of leaching of 16 inches, and a bulk density of 1.89. If the average calcium-carbonate equivalent hardness of the near-surface ground water were taken as 250 parts per million, the bulk density of the soil as 2.0, total carbonate content as 25 per cent, depth of leaching as 50 cm, with 12 cm average annual soak-in, then 11,000 years would be required for completion of this process. It may be objected that the rate of leaching may not have been constant. In fact one would expect that fewer individual rains would dissolve material as depth of leaching increases, thus slowing up the process. However, only rains that contribute to the ground water may be effective in leaching, and studies of ground-water levels indicate that such rains occur only when vegetation is at a minimum, chiefly in spring. Effect of increasing depth of leaching might then not affect the results much with such a shallow leached zone. Moreover, it is known from a study of vegetal remains, particularly in swamps, that a considerable part of postglacial time had a climate markedly warmer and drier than that of the last few thousand years. In this case the decreased use of water by vegetation might offset the reduced rainfall and permit nearly the same amount of soak-in. If there were periods of colder and moister climate than that of the present, then a longer frozen season plus reduction in chemical activity might equalize the rate of solution of carbonates. Although these determinations by weathering are admittedly rough estimates, they demand careful consideration because they were made before the radiocarbon results were made public.

Varve Age Determinations

Most geologists now agree that varves are "annual rings of the earth". Correlation of these pairs of laminae from one locality to another to obtain grand totals of age is handicapped in several ways: (1) it is difficult to connect varve records to the human calendar because varves are formed only under glacio-lacustrine environments; (2) matching of the curves of relative thicknesses is affected by the "personal equation". Very few American geologists have ever had any success in matching curves. Some who were mentioned by Antevs (1953, p. 225) as having done so did not, to the personal knowledge of Thwaites, actually accomplish anything.[5] Ebba DeGeer (1951) concluded that the radiocarbon results support the original trans-atlantic correlations of DeGeer in the Great Lakes region which many have doubted. DeGeer's correlations in Denmark agree with radiocarbon findings (Iverson, 1953). Antevs (1953), a former pupil of DeGeer, has carried on varve studies in America since 1920. Antevs

[4] Mr. F. W. Gensch of the McMahon Engineering Company of Menasha reports a discovery (Y237) in a sewer trench on Prospect Street, Menasha, in 1952, about 2 miles south of C419. He states in a letter of January 12, 1954, that "the oval shape was not caused by crushing at the time of removal. Many pieces of oval-shaped wood were found in the trench. We assume that this was caused by the weight of the ice." This specimen was determined to be 11690 ± 370 years old.

[5] Certain aspects of varve correlation do not give confidence in the results. For instance, DeGeer (1926b) correlated curves from the two hemispheres where the seasons are reversed, left gaps in curves to make them fit better (1926a) with those from other localities, ignored the effect of winds and clouds on climate, and in some places used laminated instead of varved sediments. Thwaites' surveys in Wisconsin demonstrated that DeGeer (1926a) matched curves of pre-Valders varves with curves of post-Valders sediments. Ebba H. DeGeer (1936) attempted to correlate varve curves in Sweden with the growth rings of sequoia trees in California.

(1954; 1955) concluded that the Two Creeks Forest Bed is 19,000 years old or about 8000 years older than radiocarbon indicates. He accounts for this by the hypothesis that, during decay of wood under moist conditions such as occur at the Forest Bed, more C_{12} atoms are destroyed than atoms of C_{14} and thus the latter is enriched (Antevs, 1953, p. 221–223; 1954). However, inasmuch as the proportion of the two kinds of carbon atoms is in a ratio of about a trillion to one (10^{12} to 1) obviously a very substantial portion of the wood would have to rot entirely to affect the concentration of C_{14} to a material extent. This possibility has been investigated by Armstrong (1950), Buchanan (1953), and Craig (1954). There is another possible process which apparently has been overlooked. Although the dissolved carbonates in the ground water are derived from very old carbonates in bed rock which contain almost no C_{14} the dissolving agent is carbon dioxide derived from the decay of modern vegetation which contains much more C_{14} than does the wood of the Forest Bed. Possibly some of the atoms of C_{14} in the ground water might replace C_{12} atoms in the buried vegetation. Such atomic replacement is very common in minerals. The wood and so forth of the Forest Bed has reacted with the red clay in which it is buried, bleaching it to a pale green gray for several inches from contacts. If this suspicion is correct, then radiocarbon ages of material long soaked with bicarbonate solution are not reliable. Hunt (1955) calls attention to the fact that the reasonable agreement of radiocarbon ages within a given region does not extend to other regions. It is possible to set up a radiocarbon time scale for a given region which appears correct but does not agree with that for a different region. He suggests that samples from a dry climate are more reliable than those which were preserved under moist conditions. The moist climate specimens, almost without exception give younger dates than do those of arid environments. The Forest Bed lies in such a moist region. Are all ages there from radiocarbon much too small? This almost certainly applies to specimen C-419 with an average of two measurements only about 6400 years, not much older than recorded human history. Radiocarbon dates of the Forest Bed are not much older than those of human relics in southwestern United States. Did early man live in that region when the ice front was at Milwaukee? If the dates are wrong they cannot all be ascribed to instrumental errors but imply contamination with C_{14} by some process which is not definitely known. Atomic, (not molecular) replacement from modern carbon dioxide has been suggested above. To test this theory a log from a sewer trench at Kimberly was run by Field Research Laboratory of the Magnolia Petroleum Company through the kindness of Dr. Henry Nelson. No significant age difference could be found between the core and outside showing that if contamination is present it must be uniform throughout the 7 inch log. Hunt suggests contamination by micro-organisms and possibly also by deposition of organic matter from soil solutions. No observations of micro-organisms have been recorded. It is generally thought that such are not common at a depth of more than a few inches. Organic matter is recorded in many analyses of ground water. Although most geologists in this country support the accuracy of radiocarbon dates in preference to varve chronology and estimates from leaching, melting of the ice, post-glacial erosion, and temperature changes, it is possible that the future may decide that in the Great Lakes region these are much too small, perhaps half of what they should be (Horberg, 1955, p. 281; Thwaites, 1956, p. 132).

Other Age Determinations

A common method of age determination in the past was to divide the rate per year of a given process into the total amount accomplished since glaciation. One of the most studied examples is Niagara Falls. Here both the present rate of recession and the total distance are well known. However, there have been marked variations in discharge of the river because of the use of other outlets from the upper Great Lakes. There were also changes both in height of the falls and in material beneath the falls. Students of the problem have given answers from 3000 to 50,000 years. DeGeer (1926a), using his often-questioned transatlantic varve correlations, calculated 9500 years. This result agrees reasonably well with the radiocarbon age of the Forest Bed, for the falls are almost

certainly post-Valders in age (Flint, 1953). Rate of cataract recession depends upon total fall multiplied by discharge of water per unit of width. Leopold (1953) states that mean depth of a stream of water is approximately proportional to the four-tenths power of total discharge and that velocity is related approximately to the one-third power of discharge. Hence discharge per unit of width, velocity times depth, should be approximately proportioned to the three-fourths power of discharge. If this relation holds true at Niagara Falls, erosion would be related to a power of discharge considerably less than unity. Moreover, field inspection of the Falls by Thwaites suggests that, as discharge is reduced, the width of the cataract decreases markedly. Hence it may well be that rate of erosion during low-water stages when other outlets took much of the discharge was not so much affected as some seem to have thought, and the age of the falls may be much below the maximum estimates, possibly in line with the radiocarbon age of Valders drift. This radiocarbon age does not check with the time elapsed since glaciation as calculated by Hotchkiss and Ingersol (1934), but possibly the Valders invasion was so brief that it was not registered in underground temperatures of Northern Michigan. Later investigation of the same data by Birch (1954) throws considerable doubt on the earlier calculations. Hough (1953b) attempted to date cold-water layers in cores from the deep oceans by the Urry method (Pigot and Urry, 1942; Urry, 1942; 1949); this method uses the percentage of equilibrium of uranium to ionium and radium. Hough gives results of cores in the North Atlantic Ocean, Southeast Pacific Ocean, and Ross Sea. In the first there is a thin but distinct record of a cold climate 11,000 years ago. The Pacific core records by red clay a cold period from 7000 to 12,700 years ago. Only a single pebble, which may have been ice-rafted, occupies this age level in the Ross Sea core. All three cores show the close of a long cold period at slightly less than 15,000 years before the present. These cores raise the questions: (1) did cold water in the deep oceans correspond to glacial or interglacial times on land? and (2) Is the evidence of a cold period at 11,000 years before the present a confirmation of radiocarbon dating? If the cold periods are really glacial, as most geologists think, then could the end of the long cold period at roughly 15,000 years ago confirm Antev's dating of the Valders substage? This does not seem entirely plausible, for the Valders invasion as recorded in northeastern Wisconsin was a brief event compared with the earlier Cary glaciation. Hough's conclusions therefore appear to agree with the radiocarbon method of dating for the Valders glaciation although they do not check ages of older substages of the Wisconsin (Thwaites, 1956, p. 132).

Valders Glaciation

Nomenclature

Correlation of the Valders red drift of northeastern Wisconsin with the gray Mankato drift of Minnesota is still problematical (Thwaites, 1943, p. 121; 1956, p. 86–87). It will be necessary to survey the wide gap of poorly settled country between the two localities before the term Valders can safely be dropped. The type locality for the Valders drift is the quarry at Valders where both the red and gray drifts are shown (Fig. 7).

Lobation and Direction of Flow of Valders Ice

In only a few places can the striae of the Valders invasion be separated from the older markings. At Valders the younger striae cross the Cary markings almost at right angles; they trend 260°, compared to an average of 175° for the older striae. Opinions have varied as to the direction of flow on the younger marks. Since they are best developed on the east-facing sides of minor ridges of the bedrock, movement was toward the west. At first this appears strange, for only 2 miles west at Quarry the major markings bear 180°. Weaker striae at Quarry which trend 128°–142° were found by Alden (1918, p. 317, 318) to be Valders. Alden also described markings at 218°–240° which may be of Tazewell (or older) age. The explanation seems to be that the west-trending striae at Valders are normal to the north-south boundary of the Valders red drift to the southeast (Pl. 8) and that the southeasterly striae at Quarry mark the border of a minor lobe of Valders ice controlled by the low ground along Manitowoc

River. No interlobate moraine was formed by the short-lived Valders ice. Between a locality about 2 miles east of St. Nazianz and Chilton the south border of the Valders red drift trends nearly east-west and forms a very low ridge which divides two large swamps. There are no known striae to confirm this mapping. Northwest of Chilton both the soils maps and Alden's survey indicate a marked re-entrant dividing the minor lobation (or lobations) of the Manitowoc River valley from the much larger Valders Green Bay lobe which extended south of Fond du Lac. The boundary of the red Valders till is easily distinguished not only by color of the *C* horizon of the soil profile but also by the marked reduction in number of boulders within the Valders area compared with the abundant field stone in the region not so covered. Evidently the motion of the Valders ice was locally far different from that of the Cary invasion.

Nature and Extent of Valders Till

As shown in Figure 6 the Valders till normally has a higher content of silt and clay than does the Cary till, and a corresponding reduction in larger particles. This paucity of stone content led T. C. Chamberlin to regard the entire deposit as lacustrine. The presence of striae, the fact that in many places the margin adjoins lower ground without a red till cover, and the obvious disturbance of underlying materials (Pl. 2, Fig. 1; Pl. 5) proves the glacial origin of the "red clay." Alden (1918, p. 314, 315) correctly interpreted the high clay-silt content as due to the overriding of red lake sediments deposited during the interval between Cary and Valders, the Forest Bed or Two Creeks interval as it is called today. Murray (1953) reached the same conclusion. The red till has a very patchy and irregular development because: (1) the Valders invasion could not have lasted very long; (2) the Valders till is semifluid when the water content is large and thus creeps or slides off comparatively gentle slopes. Wherever the till was thin owing either to slight accumulation or to subsequent removal by creep combined with erosion by running water, the soil-making processes destroyed its identity. This fact makes tracing of the Valders glacial margin extremely difficult and inaccurate. The scarcity of red till in Door County north of Sturgeon Bay and in considerable parts of the counties to the south has long been a puzzle. The hypothesis that these areas were nunataks is disproved both by their low elevation and by the presence of local isolated patches of red till within the areas. The suggestion that there was more than one advance of the Valders ice, thus explaining the young age of the wood (C419) from Little Lake Butte des Morts, is not in harmony with the given facts. There should then be more red till to the north instead of less. A mass of weathered red till was found in a road cut in NW ¼ SW ¼ sec. 11, T. 22 N., R. 23 E., about half a mile south of Stangleville. (Fig. 6). This till shows 37 per cent sand and 61 per cent silt and clay, much more like the red Valders till than even the most thoroughly weathered Cary gray till.

Soils on Valders Till

In the earlier soils maps the soils developed from Valders till and associated lake clays (Whitson and others, 1914; 1919; 1921) were all mapped as "Superior." In later reports (Whitson and others, 1925; 1926; 1929; Kellogg, 1930) a topographic division was followed, and the better-drained soils were classified as "Kewaunee." Marshy red-till soils and lake-clay soils have always been termed "Poygan." In general the division between Kewaunee and Superior approximates that between till and lake-clay soils, but the distinction is topographic and not textural. Because of the cool winds from Lake Michigan this area originally supported a dense forest of mixed hardwoods and conifers. Remnants of forest are now mainly confined to swampy areas. With such vegetation it follows that the soil is podzolic with a strongly bleached *A* horizon. Depth of leaching of carbonate is 15–18 inches, much less than in the Cary till.

Extensive cuts demonstrate that the Valders ice did not form any marginal moraines of its own. All important constructional land forms of the area covered by Valders till are inherited from accumulations of Cary till. Bretz (1951) demonstrated that the Port Huron Moraine of Michigan (Fig. 1), long described as the border of the "Late Wisconsin" readvance, was instead due to a readvance of the Cary ice. It appears

to correspond to the buried Cary moraine which disappears under Lake Michigan a few miles north of Port Washington. This moraine, which in Wisconsin is mantled with Valders red till, can be traced into the moraine along the lake shore in the southernmost part of the area here discussed. This moraine appears to correlate with that which extends north from Two Rivers to Sturgeon Bay, and Bretz (1951, p. 422) suggests that the Manistee Moraine of Michigan is a portion of the marginal deposit of the Valders drift. However, no ridge is found at the margin of the Valders drift in Wisconsin. Bergquist (1954) describes the color of the till in the Manistee Moraine as about the same as that of the lower or Cary till at the Two Creeks Forest Bed. Probably the Manistee Moraine is not Valders but a veneered Cary deposit, like the Port Huron Moraine; its continuation beneath Lake Michigan would bring it out to join the east side of the wide morainal belt which borders Lake Michigan (Fig. 1) north of Two Rivers. This might explain the great width of this belt of Cary accumulation. The Manistee Moraine would then be younger than the Athelstane Moraine (Thwaites, 1943, p. 133–134) which appears to join the Interlobate Moraine near Denmark.

Erosion by Valders Ice

There is little evidence of erosion by the Valders ice except in unconsolidated material (Fig. 12; Pls. 4, 5). At Valders the older Cary and perhaps some pre-Cary striae were not effaced. In many places it is highly doubtful whether all the observed striae in the Valders area are actually of that age.

In most good exposures of the Valders till there are many transported masses of silt, sand, gravel, and less abundant logs of wood all obviously derived from older deposits. Some of these transported masses are several feet in diameter. A large mass of red lake sediments was found in the delta gravel of Cary age near Manitowoc. The contact of the red till with older deposits reveals many irregularities which are ascribable to ice shove (Pl. 2, fig. 1; Pl.6). At the Two Creeks exposures of the Forest Bed the strata below the red till are much disturbed and form a series of anticlines and synclines (Fig. 11; Pl. 5, fig. 1). In some places the varved clay below the Forest Bed has been thrust up in minor anticlines displacing the entire overlying sequence up to the base of the Valders till. The Valders ice eroded the tops of some of these upthrusts (Pl. 4, fig. 2). Some of these intrusions display on a small scale the mechanism of salt-dome formation. The overturning of the trees of the Forest Bed has been ascribed to ice shove. Less than 10 good examples of such trees could be found, but they appear to show a fairly consistent direction of the fallen trunks from a direction almost due west to west-southwest (about 240°). Most of the logs which are not in position of growth are obviously driftwood in the lake sand and silt which killed the forest. At no place is there a Forest Bed where the Valders till rests directly on the gray Cary till. One tree was observed with roots in the Cary till, but it was preserved under the overlying sand and silt. As noted previously the failure to find any Forest Bed in many contacts of the Valders till on sand and silt is due to its true stratigraphic position at a lower level and not to the erosion which explains its absence on till. East of Green Bay thrust planes and faults are found in the pro-Valders sand below the red till. The strata have been contorted along these slip planes (Pl. 7, fig. 1).

Drainage of Valders Ice

Although the Valders ice invasion was brief, a considerable amount of outwash was derived during its melting. The most conspicuous line of meltwater drainage headed north of Casco and joined the valley of the West Twin River east of Larabee. The surface of this plain must originally have been about 750 feet elevation at its northernmost point, and 658 feet just north of its apparent terminus. The line of escape for the meltwater south of that point is uncertain. A low tract extends from the recognizable terrace along the course of the river southeasterly, and another along East Twin River, but subsequent erosion has obscured the relations. This line of drainage south from Casco is cut across by the valley of Kewaunee River. This valley heads on the Green Bay-Lake Michigan divide at elevation slightly above 680 feet. Below the County Park, at the old

quarry west of Kewaunee, the bottom of the river is drowned by modern Lake Michigan. This lower-level valley could not have been formed until later than the outwash train.

Lake Shoto

The lower parts of both West and East Twin River valleys show evidence of standing water here called "Lake Shoto." Evidence of a lake is found in the accompanying well log and in exposures along the Shoto-Mishicot road in NE ¼ NE ¼ sec. 25, T. 20 N., R. 24 E., where more than 10 feet of red silty clay is exposed in the edge of the area eroded by West Twin River. In SW ¼ NE ¼ sec. 28, same township, this red clay fills a hollow in older fine sand and silt of undetermined age.

Partial Log of Well of Western Condensing Co., Mishicot

Samples furnished by Layne-Northwest Co., and examined by F. T. Thwaites; elevation about 620

	Thickness, feet	*Depth, feet*
Clay, red dolomitic	5	5
Silt, gray, dolomitic	5	10
Clay, red, dolomitic	15	25
Clay, silty, gray, dolomitic	25	50
Till, pink, dolomitic	5	55
Sand, medium, gray, dolomitic	5	60
Bedrock		

Although evidence indicates the presence of a post-Valders lake, the areal extent is not known. Neither beaches nor shore features were seen in the field or on the air photographs. The tentative elevation of 650 feet as shown on the map is probably too low, for sand and gravel, possibly formed in a shallow lake, occur in the west part of sec. 6 of the same township at elevation 660. The eastern and southern borders of Lake Shoto are particularly hard to find because they may have been formed by the decaying Valders glacier. The lake seems too high to be correlated with the post-Valders-Calumet level.

Later Lake Oshkosh

Water ponded in front of the waning Valders ice in the Green Bay-Fox River lowland is termed Later Lake Oshkosh (Thwaites, 1943, p. 139–141). The first outlet was in the vicinity of Portage, Wisconsin, at an elevation of slightly above 800 feet. Evidence of this lake, in this region, consists of poorly developed beaches and terraces south of Stockbridge at an elevation not far above 800 feet. The numerous abandoned gravel pits along this beach are now well covered by vegetation. North of Stockbridge the Niagara Escarpment is too steep to have preserved any definite beach remnants, but north of Highcliff a distinct shore cliff can be traced to sec. 16, T. 20 N., R. 19 E. In NW ¼ NW ¼ sec. 15 of the same township is a narrow inconspicuous valley across the divide to Manitowoc River at elevation of about 810 feet. As soon as cleared from ice, this must have been the second outlet.

Lake Manitowoc

The Manitowoc Valley above Clarks Mills contains extensive swamps. The river has a very low slope, and low marshy ground locally extends many miles from the channel as shown on the map. When drainage from Later Lake Oshkosh escaped into this region, this became an extensive lake which the writers term "Lake Manitowoc." The beach features are not conspicuous, but Bertrand discovered gravels which by their excellent sorting and small pebbles denote beach deposits. These are located near the center of section 25, T. 20 N., R. 19 E., and near the center of section 34, T. 20 N., R. 20 E. These are on the south side of the lake as if the prevailing winds were from the north off the Valders ice. Elevations of these two localities appear to be about 810 feet. In Plate 8 the area of swamp up to slightly above 800 feet elevation has been shown as the lake bed. This ponding weakened the traces of an outlet across the divide at the head of Manitowoc River. Lake Manitowoc persisted until the valley of Manitowoc River was eroded at and below Clarks Mills, a portion of which course is in bedrock. Final extinction of the lake probably occurred when vegetation became established to form the present swamps.

Outlets

From the vicinity of Bellevue a conspicuous valley leads to West Twin or Neshota River.

This was evidently the third outlet of Later Lake Oshkosh, and, since the spillway was probably at elevation 765 feet, it must have quickly lowered the level of the lake. The divide is now much choked with swamp deposits. In SE ¼ sec. 31, T. 23 N., R. 22 E., and at other places downstream this channel was eroded down to bedrock.

Progressive recession of the ice margin uncovered lower avenues of escape south of Dyckesville (Pl. 8). Elevation of the divide here is approximately 682 feet, and hence another marked lowering of the lake took place. These outlets coalesce below and join Kewaunee River. The channels are very conspicuous on the air photographs.

Further melting of the Valders glacier opened an outlet northwest of Maplewood at an elevation of about 640 feet. This channel led to the valley of Anapee River.[6]

Draining of Later Lake Oshkosh

When Sturgeon Bay was freed of ice, the waters of Later Lake Oshkosh merged with those of Lake Algonquin in the Lake Michigan Basin. Aside from the evidence of outlet channels traces of levels of a glacial lake higher than those of Lake Algonquin are very difficult if not impossible, to discover north of the vicinity of Lake Winnebago. Possibly in this region Later Lake Oshkosh was either very short-lived or was confined to a narrow moat alongside the waning Green Bay Lobe. No trace of beaches was found by Thwaites (1943) during the survey of the area west of Green Bay.

Later Lake Chicago

Bretz (1951) and Hough (1955) have shown that the level of Lake Chicago in the Michigan basin was no higher than the Calumet or 620-foot beach of the Chicago area by the time the Valders advance took place (Fig. 13). Calumet beaches cannot be proved in this area except perhaps along Lake Shoto. Not long after the recession of the Valders began, increased supply of meltwater eroded the Chicago outlet down to bedrock at about 605 feet elevation. This initiated the Toleston stage of Lake Chicago. This outlet was to control lake levels in the Michigan basin for a long period of time. It seems probable that no shore line of the Door Peninsula area is older than Toleston.

Lake Algonquin

The name "Toleston" was applied in the Chicago area before lake records had been traced to the north. Goldthwait (1907; 1908; 1910) discovered that the name Lake Algonquin also applied to the same level. Most maps show that Lake Algonquin extended through both the Superior and Huron basins and that several outlets were used, either simultaneously or in succession. Recent studies by Hough (1953a) and Zumberge and Potzger (1956, p. 273) demonstrate that the occurrence of outwash plains in the northern peninsula of Michigan well below the highest Algonquin beach of that locality and the presence of morainal topography north of these down to 650 feet elevation prove that glacial ice remained in at least the eastern part of the Superior basin during much of the history of Lake Algonquin. This conclusion checks with the nature of the sediments of Lake Algonquin

[6] To approximate the discharge of the several outlets of Lake Oshkosh computations were made for each of the three major outlets by the use of Mannings formula; in feet per second, $v = \frac{1.5\ d^{2/3} s^{1/2}}{n}$ where v = velocity in feet per second, where d = mean depth in feet, s = slope in feet per foot, and n = roughness of the bottom. The last was assumed to be .05 which appears a fair value for a sandy bottom. Mean depth was assumed to be 10 feet, and the ⅔ power is about 4.64. Width was measured on the new topographic maps and taken at an average of 900 feet. Slopes proved difficult to determine because some of the channels have been built up by subsequent deposition in places and eroded farther downstream. The values were taken at 1 foot fall in 750 feet horizontal for the Bellevue channel, 1 foot in 1250 feet for the Dykesville channel, and 1 foot in 850 feet for the outlet leading to Anapee River. Velocities ranged from about 3.9 to about 5.0 feet per second. Discharge, Q, is velocity multiplied by width and by depth ($Q = v \cdot w \cdot d$). Results ranged from 35,400 cubic feet per second to 45,500 cubic feet per second. Agreement seems reasonable for steady flow after the channel had been graded and does not indicate initial flow. Neglecting a change in width, which would be small because of the steep banks, discharge should vary with the ⅔ power of depth.

which rest directly upon the Valders red till. Excellent exposures of varved clays with dark-red winter layers were exposed along the shore of Lake Michigan northeast of Manitowoc both during Goldthwait's survey and in the last few years. Goldthwait discovered logs of wood in these, but a thorough search has failed to find these again, nor could Alden find them in 1910 (field notes). (*See* Figure 11 for lake levels during different surveys.) Quite possibly they were washed from an adjacent exposure of the Forest Bed which is now concealed. Red Valders till appears to be present below these clays (Goldthwait, 1907, p. 59). Similar varved clays, together with sand and gravel occur at many points along the shore cliff near the Forest Bed exposures east of Two Creeks. These deposits prove that glacial ice formed a part of the shore line of the lake at this time. The varved clays of the Green Bay region, formerly excavated to make brick, were also deposited in Lake Algonquin. Lower beaches formed by the uncovering from the ice of outlets at lower elevation in Canada are shown by Goldthwait (Fig. 16) as joining the highest level south of Two Rivers, the approximate limit of the inferred area of deformation. Stanley (1937), however, shows that in Georgian Bay of Lake Huron the beaches are all parallel to the highest one and pass to the south beneath horizontal beaches of Lake Nipissing (Hough, 1955, p. 962). It is clear that in the Door Peninsula all lake features higher than the 600- to 605- foot Nipissing beach are are referable to Lake Algonquin and its immediate successors. Possibly these lower Algonquin beaches were not controlled by the rock bottom of the outlet west of Chicago but by outlets in Canada.

Low Level or Taylor Lakes

That the level of water in Lake Michigan must have receded to far below that of the present was not realized by many earlier students of the region. Stanley (1938) called attention to the submerged valley through the Straits of Mackinac (Fig. 4, 15). He had previously (1936; 1937) observed the contrast in tilting of the Georgian Bay beaches just noted above. Later mapping by Thwaites (1949) and by Emery (1951) of the bottom of Lake Michigan (Fig. 4) disclosed similar valleys in the north part of Green Bay. About the same time Thwaites received and examined cuttings from test borings for bridges at Sturgeon Bay, Kewaunee, and Two Rivers which show considerable thicknesses of marl which are unconformable on the Algonquin glacial lake clays (Fig. 14). In 1943 Thwaites also obtained cuttings from several test borings on Jones Island, Milwaukee. All this evidence combined with the results of coring in the bottom of Lake Michigan by Hough (1955) proves a very low stage of water and explains the topographic evidence of drowning shown by many of the valleys which enter the Lake Michigan basin (Fig. 14). In 1939 Antevs gave the name "Taylor Lakes" to this low level in honor of Frank B. Taylor. Hough names the low level of Lake Michigan "Lake Chippewa" (Fig. 15) and that of Lake Huron "Lake Stanley". Hough's cores show that a sand layer with shallow-water shells occurs in Lake Michigan between deeper-water sediments down to a depth of 350 feet (Hough, 1955). Laminated and varved red clays occur only below the sand layer and in the portions of the lake bottom below 350 feet depth. Some red clay above the sand layer of the shallower portions of the basin grades upward into modern gray clays. There is no suggestion of varving in this higher red clay nor is there any varved clay in the cuttings from the test holes at Milwaukee. It is entirely possible that the entire 173 feet down to bedrock on Jones Island is a filling deposited during the rise of lake level following the low stage. The only zone which suggested glacial till on first examination is from depth 116.2 to 126.3. On re-examination this was interpreted as a light-gray dolomitic or calcareous clay interbedded with fine beachlike gravel. A thin layer of red clay resting upon the bedrock could be Valders till or Algonquin clay. The bridge explorations shown in Figure 14 do not prove so much lowering of the lake level, but the marl deposits indicate warm quiet water of bays and estuaries formed during a rise of lake level.

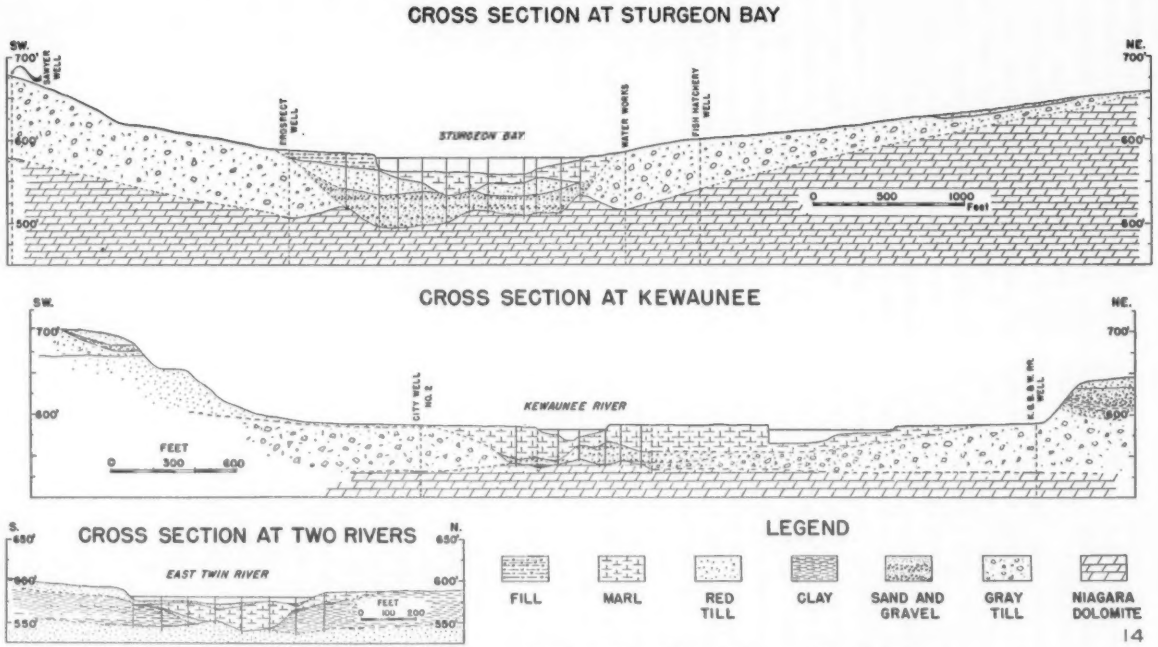

FIGURE 14.—CROSS SECTIONS AT STURGEON BAY, KEWAUNEE, AND TWO RIVERS
Showing evidence of the low stage of Lake Chippewa. Based in part on test-boring samples furnished by Wisconsin Highway Commission.

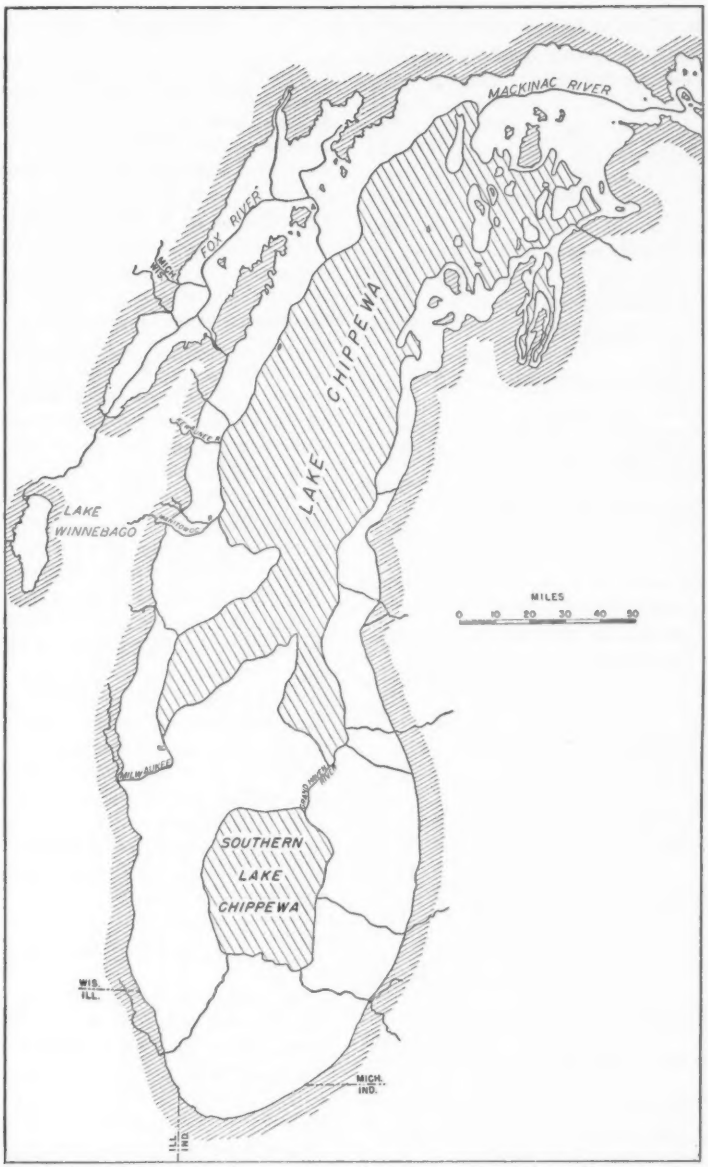

FIGURE 15.—MAP OF LAKE CHIPPEWA, THE LOW STAGE IN THE LAKE MICHIGAN BASIN BETWEEN THE ALGONQUIN AND NIPISSING STAGES

Modified from Hough with addition of streams which then ran into the lake. Only a portion of these valleys has been preserved, owing to filling by later sediments.

Log of Test Hole No. 1, Jones Island, Milwaukee

Samples furnished by Klug and Smith, Engineers, and examined by F. T. Thwaites

	Thickness, feet	*Depth, feet*
Fill and boulders, no sample	7.0	7.0
Sand, fine to gravel, fine, some silt	19.5	26.5
Marl, light to dark gray	37.4	63.9
Sand, fine, light gray, shells, wood	20.1	84.0
Sand, coarse to gravel, fine (no samples)	14.0	98.0
Sand, fine to medium, gray, some silt	5.0	103.0
Clay, silty, light gray, dolomitic	2.0	105.0
Sand, medium to fine, light gray, dolomitic	11.2	116.2
Clay, gray, dolomitic; gravel, sandy, fine	10.1	126.3
Sand, medium to very coarse, silty, gray	7.5	133.8
Sand, fine, very silty, light gray	5.5	139.3
Gravel, fine; sand, fine to very coarse, light gray	22.0	161.3
Sand, clay, small pebbles (no sample)	7.0	168.3
Clay, red, dolomitic, small stones, sand (possibly till)	5.0	173.3
Bedrock	.7	174.0

Log of Test Hole No. 5 for bridge over Kewaunee River, Kewaunee

Samples furnished by Wis. Highway Commission and examined by F. T. Thwaites

	Thickness, feet	*Depth, feet*
Marl, gray	30	30
Sand, coarse to medium, grading below to gravel, fine, some silt and clay (postglacial?)	14	44
Bedrock		

Log of Test Hole No. 10 (about 100 feet east of east end of bridge) 22nd St. Bridge over East Twin River, Two Rivers

Samples furnished by Wisconsin Highway Commission and examined by F. T. Thwaites

	Thickness, feet	*Depth, feet*
Marl, light gray	15	15
Clay, red, dolomitic (Lake Algonquin?)	29	44
Sand, medium to coarse, light gray	1	45

Log of Test Hole No. 7 for Sturgeon Bay Bridge

Samples furnished by Wisconsin Highway Commission and examined by F. T. Thwaites

	Thickness, feet	*Depth, feet*
Water	16.7	16.7
Marl, gray	13.3	30
Till, red, dolomitic (Valders)	10	40
Gravel, fine, red, mixed with clay	10	50
Gravel, fine	30	80
Bedrock		

Shore-line Delineation

The shore lines shown on the geological map (Pl. 8) reflect a Lake Chippewa stage of Lake Michigan when the major valleys were eroded to considerable depths. When the lake level rose to the Nipissing stage, deep estuaries were formed. Ahnapee River was drowned far into Door County up to the point where the 600-foot contour crosses. The East Twin River was submerged to a point above Mishicot and the West Twin River to a point above Shoto. The Kewaunee River was prevented from eroding to the depth of the others because of the rock barrier at the County Park a few miles upstream, and the Manitowoc River was also kept to a high level because of a rock ledge at Manitowoc Rapids. The upper portions of all these estuaries are now filled with alluvial debris in large part graded to the present level of Lake Michigan, but the harbors all show the effects of this period of submergence. Along large portions of the raised beaches the Algonquin and Nipissing levels are now shown along the same line because waves of the lower stage eroded the bank. In drawing the lines on the map no attempt was made to show Algonquin shores outside the later Nipissing beaches. The shore lines omit all ravines which obviously postdate their age. For instance, the Oshkosh beach line south of DePere is deeply cut by later gullies. Just how many of these gullies along Lake Michigan date from the interval between Algonquin and Nipissing is impossible to tell, and most of the smaller indentations have been omitted.

Cause of Rise in Water Level

The only cause which could bring about the rise of water level after the low stage of the Taylor lakes or Lake Chippewa is crustal movement in the region to the east (Fig. 16). It is postulated that the low outlet from Lake Nipissing through North Bay to the head of the Ottawa River then stood several hundred feet lower than it now does. Just how far west the rise extended is problematical since the bottom of Lake Huron has not yet been cored. The maximum known depth (Stanley, 1938) of the submerged valley through the Straits of Mackinac is about 150 feet at the west end and increases to slightly more than 200 feet to the

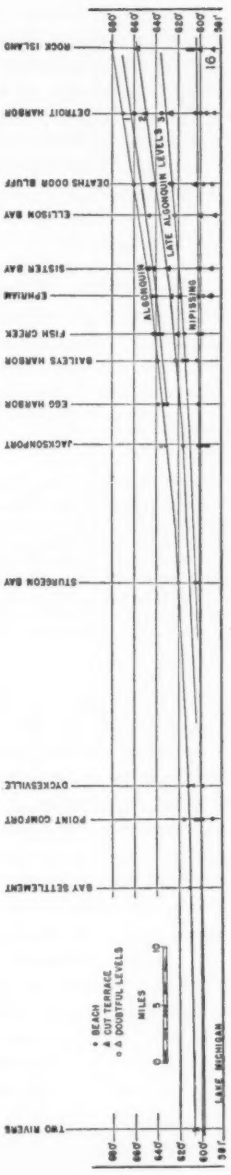

FIGURE 16.—PROFILES OF THE WARPED BEACHES OF THE DOOR PENINSULA
After Goldthwait. Note that the Nipissing beach is very little deformed

east. The amount of alteration by subsequent deposition is unknown. The outlet of the 350-foot level of Lake Chippewa has therefore been uplifted about 200 feet north of Two Rivers and 267 feet between the outlet and North Bay. If 400 miles is taken as the airline distance from the Two Rivers to North Bay, Ontario, the uplift would average about 1.1 feet per mile, less than the inclination of some of the beaches of Lake Algonquin. From the east end of the Mackinac channel to North Bay is about 250 miles, and the uplift is about 317 feet or slighly less than 1.3 feet per mile.

Correlation of the Low Stage

At first thought it might be concluded that the change between the low-water stage and the present level might have occurred at any time since Lake Algonquin, but the following facts demonstrate that it took place prior to the very well-developed Nipissing beach which is found throughout the Door Peninsula at elevation of 600–605 feet. (1) This low or "20-foot" beach is very little deformed. (2) In contrast, the Algonquin proper and lower beaches seem to be parallel and to slope strongly to the south beneath the Nipissing beach (Figure 16). (3) No varved clays are associated with Lake Nipissing, but instead several beaches, as north of Green Bay, are fossiliferous and contain snails and clams which prove a marked amelioration of climate (Leverett and Taylor, 1915, p. 432; Baker, 1920, p. 79–95; Wagner, 1905; Coleman, 1922). (4) Hough in his mimeographed report of 1953 (1953a, p. 85; 1955, p. 965) notes that on the east shore of Lake Michigan the Nipissing beach is later than drowned valleys which were eroded to a much lower level of the lake. (5) Hough in the same report (1953a, p. 90) concluded that the radiocarbon dating of wood from Dalton, Illinois, at 3469 ± 230 years and of peat from the late bottom off the Apostle Islands of Lake Superior at 3656 ± 640 years are too young for Lake Algonquin. The latest study of the late glacial and recent history of Lake Michigan is that by Zumberge and Potzger (1956). Their work was based on radiocarbon age determinations from exposures on the east side of the lake. It differed from an earlier paper by Flint (1953) but a later report by the

same author (1956) accepted their conclusions. Table 2 shows the age determinations and the concurrent vegetation shown by pollen studies.

TABLE 2—RADIOCARBON AGES OF STAGES IN LATER HISTORY OF LAKE MICHIGAN

Event with elevation of water surface	Age before present	Vegetation
Lake Michigan, 580	0	Oak-Pine
Algoma beach, 595	2,500	Oak-Pine
Lake Nipissing, 605	3,500	Oak, hemlock, hickory etc. (Thermal maximum)
Lake Chippewa, 230	5,000	Oak-Chestnut Pine declining
Lake Algonquin, 605	8,500	Spruce-Fir declining
Two Creeks, Forest Bed, ?	11,000	Spruce, Fir

Uplift of the Nipissing Beach

Within the area here described, the uplift of the Nipissing Beach is too small to measure with certainty. It has been identified to the northeast at elevation 631 feet on Mackinac Island and at 698 feet near North Bay, Ontario (Leverett and Taylor, 1915, p. 458, 459). Whatever doubts may exist as to the correct extension of the younger Algonquin beaches south of Two Rivers, there seems little question that the southern Nipissing beach is no higher than the rock sill west of Chicago. Other students (Leverett and Taylor, 1915, p. 448; Hough, 1953c; 1955) conclude that this is the same outlet that was used during one stage of Lake Algonquin. The outlet of Lake Nipissing at North Bay, Ontario, was abandoned in favor of the present one at Port Huron, Michigan, because of uplift of the land northeast of Lake Huron. The fall of water level below the Nipissing beach to the present level leaving the Chicago outlet dry is clearly due to erosion by St. Claire River below Lake Huron.

Cause of Uplift

The great majority of students (Daly, 1934; Gutenberg, 1933; Antevs, 1939) of the Great Lakes region accepted without reservation the hypothesis that the postglacial uplift of that area (Gilbert, 1898) is due to isostatic rise following melting of the ice, a process delayed by the high viscosity of the earth which retarded subsurface flow. However, not all have accepted this explanation. Taylor (Leverett and Taylor, 1915, p. 502–518) expressed doubts including (1) the relation of the hinge lines (limits of observed deformation) to probable ice thicknesses, for they seem to be parallel to the margin of the Precambrian rocks instead of to the ice border; and (2) the spasmodic timing of the uplifts. Thwaites added other objections (1956, p. 94–95). He has also noted the lack of positive proof of uplift of the beaches of Glacial Lake Oshkosh (1943, p. 139–141). Moore (1948, p. 708), who had access to all the records of modern changes in water levels observed by the U. S. Lake Survey, states:

> "In Eastern North America the line of zero movement follows rather closely the edge of the Canadian Shield. To the north the old pre-Cambrian rocks are rising, to the south the younger sedimentary rocks are subsiding. The geological record shows that the older rocks have been subject to uplift since the earliest time. The 'Postglacial Uplift,' and the movement in progress at present, may be a continuation of a movement which has been manifest since earliest times."

He also concludes that there is no connection between weight of overlying ice and direction of present change of elevation and that gravity measurements within the area show only small anomalies which average zero. It may be argued that there have been many earth movements, not only in this region but all over the world, not caused by melting ice. On the other hand, the proponents of glacial control failed to realize that the rigid upper part of the earth's crust would tend to spread the effects of ice weight over a wider area than that occupied by ice. They also seem to have failed to realize that the history of ice thicknesses and melting are but vaguely known and that the high viscosity of the subsurface materials, also little known, would cause delay in response to decrease of load. The entire matter must therefore be left with the conclusion that the explanation by glacial loading and unloading is not proved, as many geologists seem to believe.

Postglacial Weathering

Alterations of the Door Peninsula since the last glacier left the region in Algonquin time have been caused by (1) weathering, (2) erosion, and (3) sedimentation. Soils are a type of alteration in which organisms played a very important part. Reference may be made to the county soils reports and to the bulletin by Kellogg (1930) which described the profiles in the region here discussed. The effect of these alterations in destroying the color of Valders till where it was very thin have already been discussed, together with the use of rate of carbonate leaching as a time measure.

Postglacial Stream Erosion

Spectacular results of postglacial stream erosion are observed (1) along Kewaunee River and other rivers of the Lake Michigan slope, (2) on the Nigarara Escarpment where it is fronted with the terraces of sand, gravel, and red till from a short distance south of Greenleaf north to Green Bay (Figs. 9, 10), and (3) along the banks of Fox River particularly in the vicinity of Kaukauna. Evidence has already been presented to demonstrate that completion of the main parts of some of these valleys was in pre-Nipissing time since the lower parts are drowned. Gulleying reaches its maximum development in the area of pro-Valders deposits just east of Green Bay, the region described by J. W. Goldthwait (1907, p. 95) as "a belt of remarkably strong kettle moraine topography." Figure 2 of Plate 7 shows a surface view of this district. Figure 1 of Plate 7 shows a cross section of one of the prominent knobs. Goldthwait (1907, p. 94) records stream terraces along the line of Kewaunee, Green Bay, and Western Railroad in sec. 33, T. 24 N., R. 21 E., at 11, 18, and 37 feet above the creek which there has an an elevation of about 600 feet. He thought that the higher one might correspond with the Algonquin lake level which he found west of Green Bay at elevation 610. This terracing may be simply an incident in downcutting in incoherent material, for all the occurrences are close to fill in the head of Green Bay at the Nipissing level which crosses the valley outlet. Red Valders till occurs on the lower slopes, on the summits of some of the knolls, and on the sides of some higher slopes of this area. Creep might account for the position of the till on some lower slopes. The Valders ice, although it caused extensive shove of the sand, did not block any valleys, because they are younger.

The rough topography of today is clearly erosional, and Goldthwait's classification as a moraine must be rejected. As shown in Figure 2 of Plate 6 the Valders till fills kettles in the underlying stratified deposits in a manner suggesting that ice was still present in them. No red till occurs in the bottoms of the deeper valleys mapped as "Genesee loam" and "Ewen silty clay loam" soils that are on valley fill which was probably due to rise of lake level. The gentler, less-dissected area where red till is present is shown on the soils map as "Kewaunee fine sandy loam" as distinguished from the steep erosional slopes shown as "rough broken land," "Coloma fine sand," and "Ottawa loamy fine sand." Most of the erosion of the area east of Green Bay is post-Valders and very likely pre-Nipissing. South of this area, in the terrace along the escarpment between DePere and Greenleaf, extensive gulleying cuts through the Valders drift forming the miniature badlands shown in Figures 9 and 10. (This is also shown on Green Bay East, DePere, and Bellevue 7½-minute quadrangle maps.) Almost all these valleys have sides now stabilized by vegetation, and fresh exposures are rare. This higher terrace is of Cary age and is mantled by Valders till. Erosion may have occurred in the Cary-Valders interval as well as since the last ice.

Postglacial Wave Erosion

The depth of Lake Winnebago does not exceed 20 feet, and wave erosion of its banks is negligible. The level of Lake Michigan varies considerably as shown in Figure 11. Wherever the lake level is much above 581 feet above sea level, wave erosion is important in many localities. The most noteworthy is just northeast of Manitowoc (Goldthwait, 1907, p. 57–59 Pl. XI). Another important locality (Fig. 17) is from the Forest Bed exposures in sec. 11, T. 21 N., R. 24 E., north to the Manitowoc-Kewaunee county line (Amy M. Thwaites, 1931). A third is just north of the breakwater

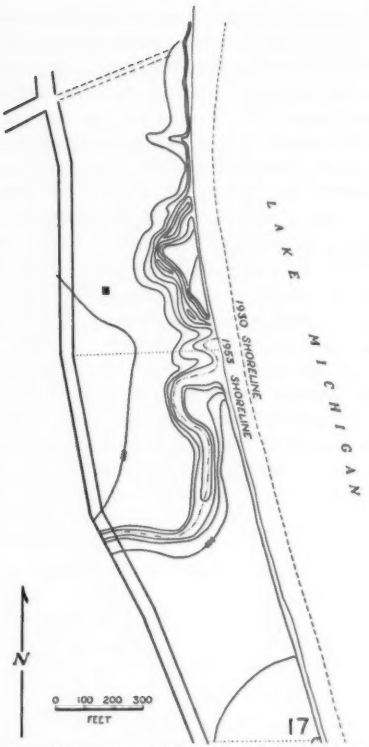

FIGURE 17.—SKETCH MAP SHOWING SHORE-LINE RECESSION AND STREAM INTERCISION
Based on paced surveys. Showing recession 1930 to 1953 and stream intercision which changed the creek outlet. (Amy M. Thwaites, 1931)

to (1) the offshore bottom profile, and (2) the disposition of eroded material by water driven toward the beach by waves. The present lake charts are not detailed enough in shallow water to be helpful. The 6-foot curve is closer to the shore near the Forest Bed than elsewhere, but this could not explain the erosion at the other localities. Hence the concentration of water driven by onshore winds may have developed its own system of debris disposal by rip currents. The building of artificial works to catch alongshore drift may, therefore, be in places more of a harm than a benefit. When there is a large amount of alongshore transporation on a beach, "starvation" of the shoreline beyond the intercepting works is recognized as a cause of increased erosion. Beach sand is known to be thin in most of, if not all, the areas of unusual erosion which suggests rapid removal. When the lake is high, it is difficult to walk along the beach on a windy day because much of the bottom is slippery clay.

Postglacial Wave Deposition

Attention has been called to the bars and barriers of the Nipissing level which were built during a rising lake level. The development of water supplies by horizontal wells beneath the lake, such as those in the glacial deltas at Manitowoc, has been discussed at Green Bay and Two Rivers. Figure 18 shows the materials penetrated in test holes on both sides of Two

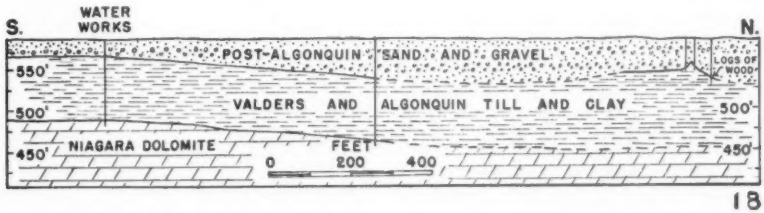

FIGURE 18.—SECTION ALONG SHORE OF LAKE MICHIGAN NEAR TWO RIVERS
Showing Nipissing sand and gravel laid down during the rise of lake level above Lake Chippewa. Data from test borings by Layne-Northwest Co., drilling contractors.

at Kewaunee. It is noteworthy that all these are in bays and not on headlands. In explaining this phenomenon, which is at variance with many textbooks, one must realize that erosion is at its maximum where material from the beach is readily removed. This relates erosion

Rivers. Two wells on Rawley Point disclosed sand with driftwood to a depth of 90 feet. Deposition was extensive during the rise of lake level before the Nipissing stage. Hough (1953a; 1955) reported several feet of gray clay under the bottom of Lake Michigan but did not

determine its calcium-magnesium ratio. In marl this is high since organisms prefer calcium for their life processes, whereas the carbonate content of the glacial clays, and of sediments washed directly from them, is dolomitic since they came from the mechanical grinding of the dolomitic bedrock. Some of the marl deposits were described by Steidtmann (1924).

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Wisconsin Geological Survey, University of Wisconsin, Madison 6, Wis.; Department of Geography, The Catholic University of America, Washington 17, D. C.

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SOME PYROXENES ASSOCIATED WITH PYROMETASOMATIC ZINC DEPOSITS IN MEXICO AND NEW MEXICO

BY VICTOR T. ALLEN AND JOSEPH J. FAHEY

ABSTRACT

Optical properties and chemical analyses are presented for 15 pyroxenes and amphiboles from zinc mines in Mexico and New Mexico. These analyses indicate that the characteristic pyroxenes in the pyrometasomatic zinc deposits examined by the writers in Mexico and New Mexico are manganoan hedenbergite and manganoan magnesian hedenbergite. At some localities johannsenite and ferroan johannsenite were formed, and later these were replaced by rhodonite coming from surges of manganese that were supplied from the invading magma of stocks, visible or concealed. Zoned birefringent garnets (andradite) were formed early in the silicate stage. On the basis of the temperature at which andradite from the Pewabic mine loses its birefringence on heating, the early stage of silicate deposition began at a temperature below 860°C. Later, at a lower temperature, possibly about 550°C., uralitization of the pyroxenes took place with a great loss of CaO and a gain in SiO_2, FeO, Al_2O_3, and MgO. The variety of uralite formed at the three mines where samples were analyzed is too low in CaO to be actinolite, and its chemical composition suggests cummingtonite. This is the first reported occurrence of cummingtonite formed by the uralitization of a calcium pyroxene. The sulfide ore minerals, sphalerite and pyrite, and also some magnetite are later than the silicate minerals and were formed by hydrothermal solutions at a somewhat lower temperature than that of the cummingtonite.

CONTENTS

TEXT

	Page
Introduction	881
Acknowledgments	883
Districts and types of deposits	883
Collection and preparation of specimens	887
Properties and composition of pyroxenes	887
Occurrence and properties of manganoan hedenbergite	889
Occurrence and properties of johannsenite	890
Occurrence and properties of rhodonite	891
Occurrence and properties of cummingtonite	892
Occurrence and properties of garnet	893
Origin and sequence of minerals	894
References cited	895

ILLUSTRATIONS

Figure	Page
1. Sketch map showing localities where pyroxenes are present	882
2. Sketch map of Central Mining district showing location of mines	884

Plate	Following page
1. Photomicrographs of cummingtonite, pyroxenes, and ore minerals	890
2. Photomicrographs of manganoan hedenbergite, garnet, and ore minerals	
3. Photomicrographs of rhodonite	
4. Photomicrographs of manganoan hedenbergite, cummingtonite, and ore minerals	

TABLES

Table	Page
1. Analyses and optical properties of pyroxenes	885
2. Analyses and optical properties of pyroxenes	886
3. Analyses of cummingtonite and pyroxenes	888

INTRODUCTION

Pyroxenes associated with the ore deposits of Mexico and New Mexico were described by Emmons (1910, p. 320) and by Lindgren, Graton, and Gordon (1910). The description of hedenbergite at Hanover, New Mexico (Lindgren, Graton, Gordon, 1910, p. 314), and the published chemical analysis of hedenbergite from the Dolores mine at Matehuala, San Luis

Potosi (Table 3, no. 6; Fig. 1), by Spurr, Garrey, and Fenner (1912, p. 468) focused attention on the nature and occurrence of the pyroxenes in these deposits. These announcements were followed by reports of hedenbergite at widely separated localities. Prescott (1916, p. 73) reported hedenbergite in the main mineral zone of the Santa Eulalia district, Chihuahua, Mexico (Fig. 1). Paige (1916, p. 15) described radiating aggregates of hedenbergite in the Fierro-Hanover district, near Silver City, New Mexico (Fig. 2). Hedenbergite was reported by Smythe (1921, p. 416) in a contact-metamorphic iron deposit near Fairview, New Mexico, by Wandke and Moore (1935, p. 774) in pyrometasomatic veins at Tepezala, Aguascalientes, Mexico, and by Loughlin and Koschmann (1942, p. 100) at the Magdalena mining district, New Mexico. This extended the reported occurrences of hedenbergite to a north-south belt exceeding 850 miles in length (Fig. 1).

FIGURE 1.—SKETCH MAP SHOWING LOCALITIES WHERE PYROXENES ARE PRESENT
□ Localities from which pyroxenes were collected
○ Additional localities from which pyroxenes have been reported

Schaller (1938) described a new manganese pyroxene, johannsenite, the manganese analogue of diopside and hedenbergite, which occurs as radiating aggregates at the Empire Zinc Company's mine near Hanover, New Mexico, at Tetela de Ocampo, Puebla, Mexico, and at Pachuca, Hidalgo, Mexico. Schmitt

(1939, p. 791–795) published descriptions and analyses of salite, a member of the diopside-hedenbergite series, from the Pewabic mine, Hanover, New Mexico. These papers established that more than one species of pyroxene occurred as radiating aggregates throughout this region.

At the annual meeting of The Geological Society of America at Detroit, Allen and Fahey (1952) described a new occurrence of salite at Iron Mountain, Missouri, and Harrison Schmitt encouraged them to apply for a grant to undertake a study of the pyroxenes associated with the pyrometasomatic zinc deposits of New Mexico and Mexico.

During the summer of 1952, Allen collected 105 specimens from 17 mines in New Mexico and Mexico. These were studied by petrographic methods in thin sections and by oil immersion. Fifteen of these were selected for chemical analysis and were carefully prepared by hand picking and by using heavy solutions and the electromagnet. Eleven complete new chemical analyses and four partial analyses were made by Fahey in the laboratories of the U. S. Geological Survey. This paper is based on this material.

Acknowledgments

Field collecting by Allen during the summer of 1952 was financed by a project grant from the Penrose Bequest of The Geological Society of America. Dr. Harrison Schmitt, consulting mining geologist of Silver City, New Mexico, made available his detailed knowledge of the district, offered suggestions on mineral localities, and extended many courtesies during the field work. Mr. Don Purvis, Superintendent, and Mr. Bert Morrison, Geologist, of the American Smelting and Refining Company, Mr. J. T. Lewis, Superintendent, and Mr. F. S. Cook, Geologist, of the U. S. Smelting and Refining Company, co-operated in the collecting of specimens from the mines near Vanadium, New Mexico. Collecting in Mexico was facilitated by the courtesies of officials of the American Smelting and Refining Company; Mr. C. F. Jordan, El Paso, Texas; Mr. William P. Hewitt, Geologist, San Antonio mine, Santa Eulalia district, Chihuahua; Mr. W. J. Nock, Superintendent and Mr. Edwin D. Bieber, Mining Engineer, Hidalgo and Tecolotes mines, Santa Barbara, Mexico. The X-ray patterns of the pyroxenes and rhodonite were made by Fred A. Hildebrand, and the spectrograms were made by Harry Rose in the laboratories of the U. S. Geological Survey. Robert Reynolds co-operated by checking the indices of refraction of the analyzed pyroxenes, using a single variation method. Research equipment that was purchased for this study was made possible by a grant from the Research Corporation of New York.

Districts and Types of Deposits

The districts from which specimens were collected for description in this paper are the Central mining district, the Magdalena district, the Organ Mountain district in New Mexico, the Santa Eulalia district and the Santa Barbara district, Chihuahua, Mexico (Fig. 1).

The Central mining district (Fig. 2) is in Grant County about 10 miles east of Silver City, New Mexico. Parts of this area have been mapped and described by Lindgren, Graton, and Gordon (1910), by Paige (1916), by Spencer and Paige (1935), by Lasky (1936), and by Orodonez, Baltosser, and Martin (1955). This district contains several stocklike bodies of granodiorite and allied rocks, around which are clustered the productive mines. The Hanover pluton is known from the excellent paper of Schmitt (1939) on the Pewabic mine, a paper by Schaller (1938) on johannsenite from the Empire Zinc Company's mine at Hanover, and earlier papers by Paige (1916), Spencer and Paige (1935), and by others. At the northern end of this pluton is the Fierro mine, and to the west is the Pearson mine. A stock containing the important disseminated copper deposit that is mined by the Kennecott Copper Corporation in a large open pit occurs near the village of Santa Rita (Spencer and Paige, 1935; Kerr *et al.*, 1950). To the north of this stock are the Kearny, Oswaldo No. 1, and Oswaldo No. 2 mines. A small intrusive body forms the locus of the Copper Flat mine, which first attracted interest as a copper prospect (Lindgren, Graton, and Gordon, 1910) and

during World War II was operated as a zinc mine by the Peru Mining Company. Mineralization related to the intrusive bodies of the Ground Hog mine was not recognized until

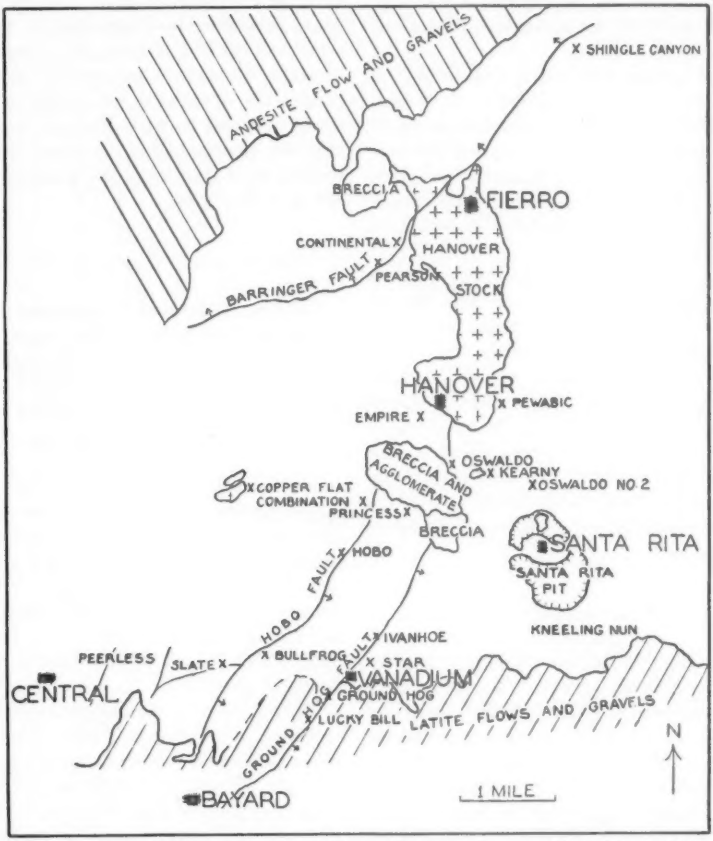

FIGURE 2.—SKETCH MAP OF CENTRAL MINING DISTRICT SHOWING LOCATION OF MINES

later (Spencer and Paige, 1935; Lasky, 1936), but it is to these intrusions that the deposits of the Star, Hobo, and Princess mines are related. The mineral assemblages of the zinc deposits of the Central mining district and their occurrence in or near limestones in close association with these intrusions indicate that they are typical pyrometasomatic ore deposits.

The Magdalena mining district in Socorro County, New Mexico (Fig. 1), was described by Lindgren, Graton, and Gordon (1910, p. 246) in a discussion of the ore deposits of New Mexico. Laughlin and Koschmann (1942, p. 100) published a detailed description of the mines of this district and recorded the optical properties of a variety of hedenbergite that gave a test for manganese. The specimen of manganoan hedenbergite that was used for the analysis (Table 2, no. 1) came from the Linchburg Tunnel in the Magdalena district and is associated with sphalerite, garnet, magnetite, chalcopyrite, pyrite, calcite, and quartz (Pl. 2, fig. 4), a pyrometasomatic mineral assemblage found in the mineralized Kelly Limestone.

The Organ Mountain district is in Dona Ana County, New Mexico (Fig. 1). The Quickstrike mine, N ½ sec.32, T.21 S., R.4 E., is located in a pegmatite (Dunham, 1935, p. 113). The ferroan

TABLE 1.—ANALYSES AND OPTICAL PROPERTIES OF SOME PYROXENES

	1	2	3	4	5	6	7	8	9
SiO_2	53.62	52.20	48.52	....	48.98	48.94	....	48.15	48.62
Fe_2O_3	0.09	0.74	1.86	0.97	1.07	0.93	0.20	1.58	
FeO	6.51	14.84	18.56	9.70	11.30	10.68	2.58	4.28	....
Al_2O_3	0.50	0.26	1.67	....	0.84	0.48	....	0.31	....
TiO_2	0.06	0.01	0.08	....	0.14	0.10	....	....	....
MnO	1.78	2.17	2.66	10.90	14.13	15.53	25.90	21.37	28.69
CaO	23.78	22.19	21.71	....	20.64	21.10	....	20.88	22.69
MgO	13.50	7.50	4.11	....	2.19	2.36	....	0.48	....
Na_2O	0.27	nd*	nd	....	0.10	nd	....	nd	....
K_2O	0.09	nd	nd	....	0.05	nd	....	nd	....
H_2O	0.20	0.38	0.99	....	0.44	nd	....	1.20	....
CO_2	....	....	....	....	....	....	....	1.39	....
ZnS	....	....	....	....	....	....	....	0.17	....
Total	100.40	100.29	100.16	....	99.88	100.12	....	99.81	100.00
α	1.685	1.687	1.690	1.695	1.703	1.708	1.708	1.713	1.710
β	1.691	1.692	1.693	1.702	1.711	1.718	1.718	1.721	1.719
γ	1.711	1.721	1.715	1.723	1.732	1.736	1.737	1.740	1.738
$\gamma - \alpha$	.026	.034	.025	.028	.029	.028	.029	.027	.028
Sign	+	+	+	+	+	+	+	+	+
$Z \wedge C$	40°	54°	38°	46°	46°	47°	46°	48°	....
2V	56°±	60°±	60°±	60°±	70°±	70°±	70°±	70°±	....
H (per cent)	20	51	67	32	38	33	9	16	....
J (per cent)	5	7	9	38	49	53	90	81	100
D (per cent)	75	42	24	30	13	14	1	3	....
Color in hand specimen	Colorless to dark greenish gray	lincoln green	slate olive	pale olive	dark yellowish brown	light olive gray	pale brown		

H = Hedenbergite = $FeO \cdot CaO \cdot 2SiO_2$
J = Johannsenite = $MnO \cdot CaO \cdot 2SiO_2$
D = Diopside = $MgO \cdot CaO \cdot 2SiO_2$

1. Ferroan diopside (salite) from pegmatite at Quickstrike mine, N ½ sec. 32, T. 21 S., R.4 E. Organ Mountains, Dona Ana County, New Mexico. J. J. Fahey, analyst

2, 3. Manganoan magnesian hedenbergite (salite), Pewabic mine, Hanover, New Mexico. Chemical and optical data after Schmitt (1939). R. B. Ellestad and A. Willman, analysts

4, 5, 6. Ferroan johannsenite, Star mine, Vanadium, New Mexico, J. J. Fahey, analyst.

7. Johannsenite, Hobo mine, Vanadium, New Mexico. J. J. Fahey, analyst

8. Johannsenite, Empire Zinc Company mine, Hanover, New Mexico (Schaller, 1938, p. 579). W. T. Schaller, analyst

9. Theoretical composition and optical properties of pure johannsenite, $CaMnSi_2O_6$ (Schaller, 1938, p. 581)

diopside listed in Table 1, no. 1, is from a crystal over 6 inches long. It is associated with quartz crystals a foot long, orthoclase, epidote, chlorite, pyrite, sphalerite, chalcopyrite, galena, and argentite. This is not the mineral association that characterizes the typical pyrometasomatic zinc deposits of New Mexico, and the features of this deposit suggest a pegmatite.

The Santa Eulalia district is located in Chihuahua, Mexico (Prescott, 1916). The specimens from this district that were used in this study came from the San Antonio mine

TABLE 2.—ANALYSES AND OPTICAL PROPERTIES OF SOME PYROXENES

	1	2	3	4	5	6	7	8
SiO_2	48.78	48.90	48.60	49.58	47.66	....	....	48.4
Fe_2O_3	0.22	0.64	0.53	4.30	2.96	0.42	1.32	....
FeO	18.18	20.32	20.64	16.93	18.72	19.47	20.08	29.4
Al_2O_3	0.90	1.96	0.78	0.38	0.70	....	....	....
TiO_2	0.04	0.10	0.10	0.06	0.06	....	....	....
MnO	8.51	4.43	5.07	4.62	4.30	4.85	4.58	....
CaO	21.14	21.20	21.22	20.46	21.02	....	....	22.2
MgO	1.26	2.21	1.98	3.35	2.48	....	....	....
Na_2O	0.21	0.25	0.21	0.30	0.26	....	....	....
K_2O	0.14	0.09	0.12	0.15	0.10	....	....	....
H_2O	0.38	0.38	0.46	0.42	2.16	....	....	....
CO_2	0.08	....	0.20	....	....	....	....	....
Total	99.84	100.48	99.91	100.55	100.42	....	....	100.0
α	1.719	1.722	1.723	1.725	1.723	1.722	1.725	1.732
β	1.730	1.729	1.729	1.736	1.731	1.730	1.732	1.740
γ	1.750	1.751	1.750	1.754	1.753	1.749	1.750	1.757
$\gamma - \alpha$	.031	.029	.027	.029	.030	.027	.025	.025
Sign	+	+	+	+	+	+	+	+
$Z \wedge C$	48°	41°	42°	47°	47°	47°	47°	47°
2V	68°±	60°±	60°±	70°±	68°±	68°±	66°±	70°±
H (per cent)	62	72	71	64	70	65	72	100
J (per cent)	30	15	17	16	15	16	15	....
D (per cent)	8	13	12	20	15	19	13	
Color in hand specimen	Grayish green	dark greenish gray	greenish black	greenish black	greenish black	greenish black	greenish black	

H = Hedenbergite = $FeO \cdot CaO \cdot 2SiO_2$
J = Johannsenite = $MnO \cdot CaO \cdot 2SiO_2$
D = Diopside = $MgO \cdot CaO \cdot 2SiO_2$

1. Manganoan hedenbergite, Linchburg tunnel, Magdelena, New Mexico. J. J. Fahey, analyst
2. Manganoan hedenbergite, Oswaldo No. 2 mine, Santa Rita, New Mexico. J. J. Fahey, analyst
3. Manganoan hedenbergite, Oswaldo No. 2 mine, Santa Rita, New Mexico. J. J. Fahey, analyst
4. Manganoan magnesian hedenbergite, 9th level, San Antonio mine, Santa Eulalia, Chihuahua, Mexico. J. J. Fahey, analyst
5, 7. Manganoan hedenbergite, 1400-foot level, Hidalgo mine, Santa Barbara, Mexico. J. J. Fahey, analyst
6. Manganoan magnesian hedenbergite, 1400-foot level, Hidalgo mine, Santa Barbara, Mexico. J. J. Fahey, analyst
8. Theoretical composition and optical properties of pure hedenbergite, $CaFeSi_2O_6$

(Fig. 1), 25 km southeast of Chihuahua City. The San Antonio deposits that contain zinc ores, pyroxenes, amphiboles, garnets, magnetite, hematite, and sulfides are considered by Hewitt (1943, p. 201) to be metasomatic replacements of limestones and best classed in the pyrometasomatic group.

The Santa Barbara group of mines are situated about 12 miles southwest of Parral in southern Chihuahua, Mexico (Fig. 1). The specimens of manganoan hedenbergite listed in Table 2, nos. 5, 6, 7, are from the 1400-foot level of the Hidalgo mine. These were collected at intervals of about 200 yards to determine any variation in the pyroxenes. Specimens from the 1200-foot level of the La Seca-Coyote vein in the Tecolotes mine have optic properties similar to those from the Hidalgo mine. Schmitt

(1928; 1930) considers that these veins were formed chiefly by replacement of rock along fissures. Birefringent garnet has been observed with the pyroxenes in some of the sphalerite ores, and this assemblage suggests pyrometasomatic replacement deposits.

Collection and Preparation of Specimens

During the collection of samples in the field, an effort was made to select only specimens of coarse grain that were free from stains and inclusions and from which could be separated uncontaminated grains of pyroxenes or amphiboles. Large pieces of ore and gangue minerals were collected from fresh faces in the mines, from the cars ready for shipment, or from the conveyors bringing the product of the day's mining to the surface. Specimens in which the long blades of pyroxene were stained with iron or manganese oxides or specimens that contained visible inclusions of sulfides or other minerals were rejected. The specimens selected each averaged about 4 pounds, but many choice specimens were larger. An effort was also made to select all the colors, shades, and variations in crystal habit, so that all the varieties of pyroxenes available at each locality would be represented.

In the laboratory, each specimen was examined in thin section and in oils by the immersion method. Each specimen that offered promise of yielding a homogeneous sample of pyroxene or amphibole was broken into small chips. Chips that were stained or showed more than one mineral were rejected. Visibly uncontaminated chips were crushed to pass an 80-mesh screen, and the portion less than 250 mesh was discarded. Grains smaller than 80 mesh were examined under a binocular microscope to observe that each grain was free of stains, inclusions, or attached grains. If only a few grains were stained, they were removed by hand picking, but if many grains were stained, the sample was rejected. Magnetite or grains including a magnetic oxide and iron fragments from grinding were removed with an alnico magnet. Calcite and other soluble carbonates were removed by immersion of the grains in cold dilute HCl (1:6) for 1 hour. This treatment did not affect the pyroxenes or the amphiboles, because the filtrate tested contained no iron, and the optical properties of the grains measured were the same after as before acid treatment. Quartz grains that escaped removal during the preliminary examination were removed by rejecting the part of each screen size that floated in a solution with a density of 2.9. Sulfide ore minerals attached to grains were removed by rejecting the part of the sample that would sink in a solution heavier than that of the pyroxene or amphibole that was being purified.

This method was more successful in obtaining uncontaminated samples of pyroxenes than of amphiboles. Samples 1, 2, and 3 (Table 3) contain less than 1 per cent visible quartz, but the high SiO_2 of the chemical analysis suggests that some imperceptible free silica is present along the fibers of the amphibole. Specimen 2 (Table 3) contains 1.84 per cent CO_2, which suggests that some undissolved calcium magnesium carbonate remains between the fibers of the amphibole.

Properties and Composition of the Pyroxenes

In the hand specimen the pyroxenes listed in Tables 1 and 2 are light olive gray, pale brown, dark yellow brown, dark greenish gray, pale olive, and greenish black, according to the Rock Color Chart (National Research Council, 1948). On the outcrop or mine dump, the color of the pyroxenes changes to a bronze that penetrates inward with increasing period of exposure. Black stains of the oxide of manganese, and red, yellowish, or brownish stains of the oxides of iron coat the surface of specimens in which the sulfide minerals and the pyroxenes have been oxidized for long periods.

The color of these pyroxenes does not seem to bear a definite relation to chemical composition or to the species of pyroxene. Dunham (1935, p. 113) described the pyroxene at the Quickstrike mine as dark green and as occurring as bladed crystals up to 16 cm in length, with $\alpha = 1.690, \beta = 1.695, \gamma = 1.713$. On the basis of these indices of refraction, he estimated from Winchell's curve that the pyroxene is equivalent to 73 per cent diopside and 27 per cent hedenbergite. Dunham did not know that some manganese was present in the pyroxene. With this exception, his determination of the optical

TABLE 3.—ANALYSES OF CUMMINGTONITE AND PYROXENES

	1	2	3	4	5	6	7	8
SiO_2	60.42	64.72	74.18	51.09	57.26	43.26	50.97	48.64
Fe_2O_3	1.24	0.88	0.35	....	1.73	1.73	7.08	1.16
FeO	22.91	9.00	5.40	32.07	15.64	23.11	6.96	15.45
Al_2O_3	0.78	0.36	0.36	0.95	0.75	3.01	0.95	0.88
TiO_2	0.06	0.02	0.04	nd	nd	nd	0.20	nd
MnO	2.82	1.65	1.01	1.50	....	0.66	....	10.77
CaO	2.82	5.54	3.86	tr	tr	19.93	20.96	21.99
MgO	4.45	11.28	12.12	10.29	21.70	1.25	11.55	0.03
Na_2O	0.22	0.15	0.14	0.75	2.80	0.21	1.86	nd
K_2O	0.31	0.10	0.10	tr	tr	0.10	0.05	nd
H_2O	3.38	2.88	2.12	3.04	....	nd	0.19	1.05
CO_2	nd	1.84	0.10	....	....	....	....	0.14
ZnO	0.32	....	....	....	....	2.15[6]		0.24
Total	99.73	98.42	99.78	99.69	99.88	95.41	100.77	100.35
α	1.623	1.617	1.602	....	....	....	1.700	....
β	1.644	1.633	1.618	....	....	....	1.711	....
γ	1.655	1.643	1.628	....	....	....	1.728	....
$\gamma - \alpha$	.032	.026	.026	....	....	....	.028	....
Sign	—	—	—	....	....	....	....	....
$Z \wedge C$	8°	13°	16°	....	....	....	51°	
2V	72°±	75°±	74°±	....	....	....	77°±	....
XY	pale brown	colorless	colorless	....	....	....	....	....
Z	light brown	pale brown	pale green	....	....	....	....	....
Color in hand specimen	olive gray	light greenish gray	yellowish gray	ash gray	....	....	....	....

tr = Trace

[6] FeS_2

1. Cummingtonite, 10th level, San Antonio mine, Santa Eulalia, Chihuahua, Mexico. J. J. Fahey, analyst

2. Cummingtonite, Pearson mine, Fierro-Hanover district, Grant County, New Mexico. J. J. Fahey, analyst

3. Cummingtonite, Kearny mine, Fierro-Hanover district, Grant County, New Mexico. J. J. Fahey, analyst

4. Cummingtonite, Cummington, Massachusetts (Smith and Brush, 1853). H. Muir, analyst

5. Cummingtonite, Baltimore, Maryland (Palmer, 1885). C. S. Palmer, analyst

6. Hedenbergite, Dolores mine, Matehuala, San Luis Potosi, Mexico (Spurr, Garrey, and Fenner, 1912, p. 468). C. N. Fenner, analyst

7. Augite, Cerro Mercado iron deposit, Durango, Mexico (Hess, 1949, p. 664). W. F. Foshag, analyst

8. Manganoan hedenbergite, Fierro-Hanover district, Grant County, New Mexico. W. T. Schaller, analyst (*in* Wells, 1947, p. 104)

properties, his estimate of the molecular composition, and the dimensions of his crystals are similar to those of the pyroxene (Table 1, no. 1) collected by Allen from the same tunnel several years later. However, the writers' specimen (Table 1, no. 1) is colorless to dark greenish gray. Thus, the color of the pyroxene probably varies in different parts of the same pegmatite.

Schmitt (1939, p. 794) records the pyroxene from the Pewabic mine, Hanover, New Mexico

(Table 1, no. 2), as lincoln green and that of no. 3 (Table 1) as deep slate green. The pyroxene with 38 per cent johannsenite (Table 1, no. 4) is pale olive, one with 49 per cent johannsenite (Table 1, no. 5) is dark yellowish brown, one with 53 per cent johannsenite (Table 1, no. 6) is light olive gray, and one with 90 per cent johannsenite (Table 1, no. 7) is pale brown. The pyroxenes listed in Table 2 range from grayish green through dark greenish gray to several shades of greenish black, with a variation in the theoretical hedenbergite molecule from 62 to 72 per cent. The color does not seem to help in estimating the amount of each molecule that is present in these pyroxenes.

The pleochroism of these pyroxenes is not perceptible in thin sections of normal thickness. Some have a faint pleochroism that can be seen in thick grains, but pleochroism is not a characteristic property on which a reliable determination of the variety of pyroxene can be based.

Schaller has prepared curves for the diopside-hedenbergite series and for a diopside-johannsenite-hedenbergite triangular diagram that have been published by Kennedy (1947, p. 565–566). During this investigation, optical methods were sought that would enable one to decide if the increase in the indices of refraction of these pyroxenes was connected with increase of manganese or with increase of iron. The iron end member, hedenbergite, can be separated from the manganese end member, johannsenite, on the basis of optical properties. However, all the pyroxenes from the pyrometasomatic zinc deposits of New Mexico and Mexico that were analyzed chemically contain considerable amounts of manganese, and thus far optical methods of determining the amount of manganese in the intermediate members of the series have not been found. Hess (1949, p. 641) published curves with salite and ferrosalite as intermediate members of the diopside-hedenbergite series that are useful for pyroxenes low in manganese belonging to this series. According to Schaller's triangular diagram (Kennedy, 1947, p. 566), however, increase in index of refraction is related to increase of manganese and to increase of iron in various combinations and proportions. Therefore, wherever possible it is desirable to indicate the dominant molecule in naming these pyroxenes and to express the presence of the subordinate element by adjectives as suggested by Schaller (1930, p. 571). On this basis four types of pyroxenes occur in association with the pyrometasomatic zinc deposits of New Mexico and Mexico, manganoan hedenbergite, manganoan magnesian hedenbergite, johannsenite, and ferroan johannsenite (Tables 1, 2).

Occurrence and Properties of Manganoan Hedenbergite

Spencer and Paige (1935, p. 59) observed that hedenbergite usually occurs with zinc sulfide, characteristically in places somewhat more distant from the intrusive mass than are garnet and epidote in the Central mining district. Schmitt (1939, p. 795, 807) noted that in the metamorphic aureole at the Peru Mining Company and Empire Zinc Company properties coarse salite rarely occurs without associated sphalerite. Thus, pyroxenes were recognized early as guides to the ore. The writers' experience confirms these observations, and many specimens of pyroxene were rejected for chemical analyses because they contained many inclusions of sphalerite grains from which they could not be satisfactorily separated. Typical occurrences of sphalerite in manganoan hedenbergite are shown at Oswaldo No. 1 mine, Santa Rita, New Mexico (Pl. 1, fig. 3), and at the Star mine, Vanadium, New Mexico (Pl. 4, fig. 5). Manganoan hedenbergite, free from sulfides and with its characteristic bladed or columnar structure, is shown from Chihuahua, Mexico (Pl. 4, fig. 1), from Santa Barbara, Mexico (Pl. 4, fig. 4), from the Linchburg Tunnel, Magdalena, New Mexico (Pl. 2, fig. 1), and from the Oswaldo No. 2 mine, Santa Rita, New Mexico.

In the analyzed samples of manganoan hedenbergite and manganoan magnesian hedenbergite (Tables 1, 2), the hedenbergite molecule ranges from 51 per cent (Table 1, no. 2) to 72 per cent, and the johannsenite molecule from 7 to 30 per cent. On the basis of this study, the average specimen of manganoan hedenbergite found in association with the pyrometasomatic zinc deposits of New Mexico and Mexico has a composition closer to that of the slate-olive variety of Schmitt (Table 1, no. 3) than to that of his lincoln-green variety (Table 1, no. 2). Unfortunately, all the specimens of pyroxenes

with a mean index of refraction less than 1.695 collected during this investigation contained needles of an amphibole and were rejected for analysis.

Koch (1956, p. 22) reports abundant diopside in the lower levels of the Frisco mine, in the Parral district of southern Chihuahua, Mexico. He gives no optical data and no chemical analysis, and the manganese content of his specimens is unknown. He states that the composition varies but is close to that of the variety salite, with a ratio Mg:Fe of about 1:1.

Occurrence and Properties of Johannsenite and Ferroan Johannsenite

The occurrence of johannsenite and ferroan johannsenite is probably similar to that of manganoan hedenbergite, because no consistent variation in occurrence has been demonstrated. The difficulty of restricting the occurrence stems from the lack of a reliable field method of identifying the various manganese pyroxenes. All give a borax-bead test for manganese; their colors, weathering, and increase in indices of refraction are not limited to any particular pyroxene. At the mines near Vanadium, New Mexico, the pyroxenes in association with rhodonite have the composition of johannsenite (Table 1, no. 7) or ferroan johannsenite (Table 1, nos. 4, 5, 6; Pl. 3, figs. 1–6). This suggests that rhodonite may be an indicator of the proximity of johannsenite or ferroan johannsenite.

At the Star and Hobo mines, Vanadium, New Mexico, specimens of the pyroxene that occur in proximity to rhodonite were selected for chemical analysis. After separating by hand-picking and using heavy solutions, the samples were studied by X-ray methods, and Fred A. Hildebrand of the U. S. Geological Survey detected no impurities in appreciable quantities. The chemical analysis of no. 4 (Table 1) contains 10.9 per cent MnO corresponding to 38 per cent johannsenite; no. 5 (Table 1) contains 14.13 per cent MnO or 49 per cent johannsenite; no. 6 contains 15.53 per cent MnO or 53 per cent johannsenite; no. 7 (Table 1) contains 25.9 per cent MnO or 90 per cent

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Plate 1.—PHOTOMICROGRAPHS OF CUMMINGTONITE, PYROXENES, AND ORE MINERALS

Figure 1.—Manganoan hedenbergite (H) shows typical bladed structure. Oswaldo No. 2 mine, Santa Rita, New Mexico. (*See* Table 2, no. 3, for analysis.)

Figure 2.—Separated areas of manganoan hedenbergite (H) extinguish at the same time, suggesting remnants of one large crystal. Cummingtonite separates and replaces manganoan hedenbergite. Magnetite (M) and pyrite (PR) replace cummingtonite. Star mine, Vanadium, New Mexico.

Figure 3.—Sphalerite (S), magnetite (M), and pyrite (PR) replace manganoan hedenbergite (H). Oswaldo No. 1 mine, Santa Rita, New Mexico.

Figure 4.—Cummingtonite (CU) with typical fibrous structure cut by sphalerite (S) and magnetite (M). Kearny mine, Hanover district, New Mexico. (*See* Table 3, no. 3, for analysis.)

Figure 5.—Cummingtonite (CU) replaced by quartz (Q), calcite (C), and magnetite (M). Empire Zinc Company mine, Hanover, New Mexico.

Plate 2.—PHOTOMICROGRAPHS OF MANGANOAN HEDENBERGITE, GARNET, AND ORE MINERALS

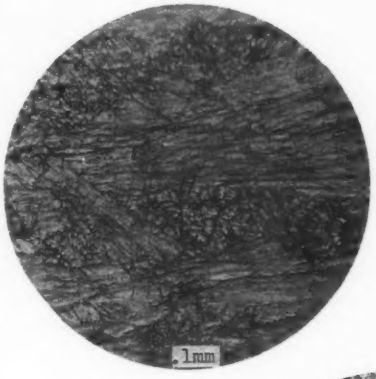
Figure 1.—Manganoan hedenbergite (H) shows bladed structure. Used for analysis (Table 2, no. 1). Linchburg Tunnel, Magdalena, New Mexico.

Figure 2.—Manganoan hedenbergite (H) is altered to fibrous cummingtonite (CU) and cut by sphalerite (S) and magnetite (M). Quartz (Q) near center contains needles of cummingtonite. Star mine, Vanadium, New Mexico.

Figure 3.—Zoned garnet (G) is replaced by manganoan hedenbergite (H) and calcite (C). Princess mine, Vanadium, New Mexico.

Figure 4.—Zoned garnet (G) is replaced by cummingtonite (CU) which in turn is replaced by quartz (Q), calcite (C), sphalerite (S), and pyrite (PR). Quartz (Q) contains needles of cummingtonite that do not show in this section. Linchburg Tunnel, Magdalena, New Mexico.

Figure 5.—Zoned garnet (G). In other parts of the slide, garnet is cut by cummingtonite which in turn is cut by pyrite, magnetite, and sphalerite. Pearson mine, near Fierro, New Mexico.



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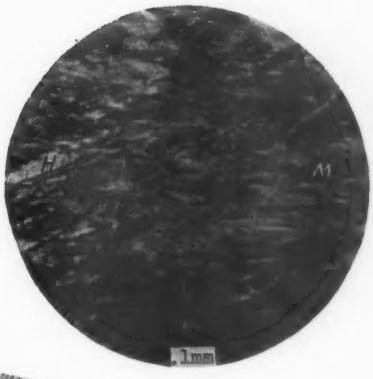

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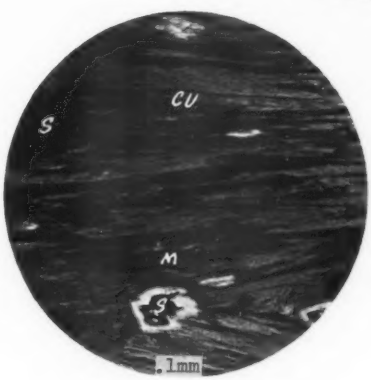

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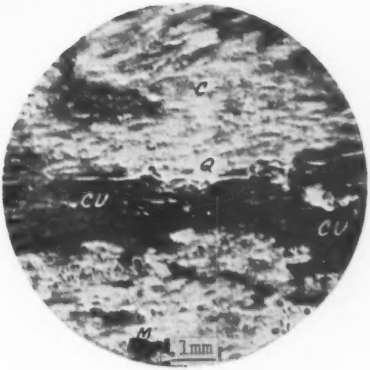

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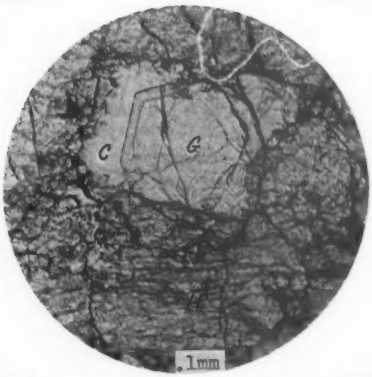
PHOTOMICROGRAPHS OF CUMMINGTONITE, PYROXENES, AND ORE MINERALS

1

2



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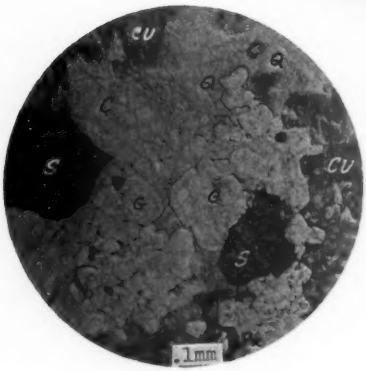

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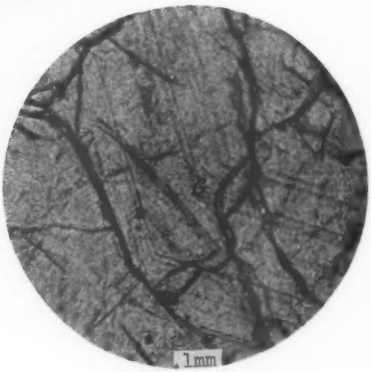

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PHOTOMICROGRAPHS OF MANGANOAN HEDENBERGITE, GARNET, AND ORE MINERALS

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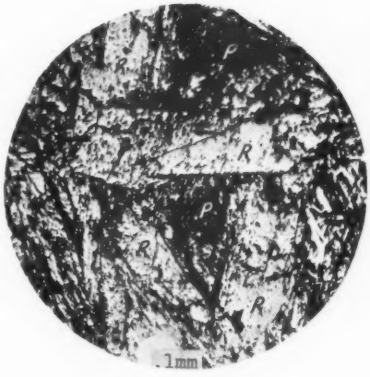

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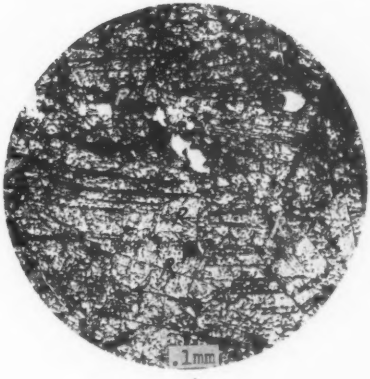

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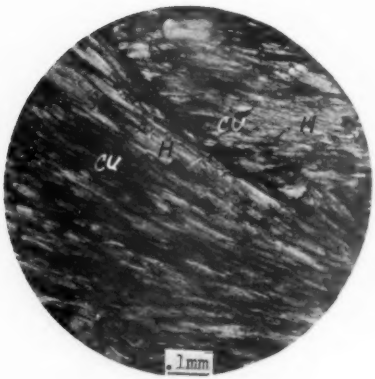
PHOTOMICROGRAPHS OF RHODONITE

1

2



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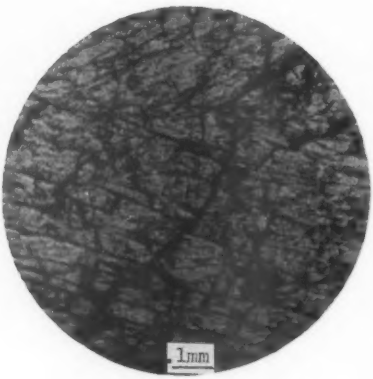

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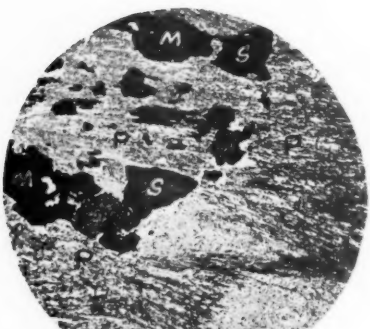

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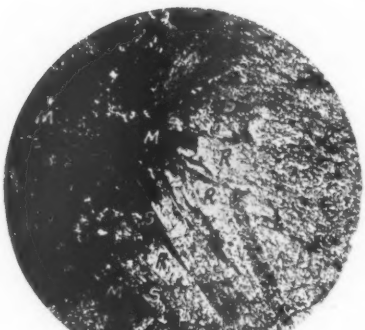

6

PHOTOMICROGRAPHS OF MANGANOAN HEDENBERGITE, CUMMINGTONITE, AND ORE MINERALS

johannsenite. The composition and optical properties of johannsenite from the Empire Zinc Company mine, Hanover, New Mexico, by Schaller (Table 1, no. 8) and the theoretical composition of johannsenite (Table 1, no. 9) are given for comparison.

Occurrence and Properties of Rhodonite

In the deep levels of the mines near Vanadium, New Mexico, a pink or flesh-colored silicate was observed that had not been recorded at the Ground Hog, Star, Hobo, and Princess mines (Northrop, 1944). Optical, spectrographic, chemical, and X-ray data indicate that this mineral is rhodonite with $\alpha = 1.717; \beta = 1.721; \gamma = 1.729; \gamma\text{-}\alpha = 0.012$; $2V = 76°$; positive; $Z \wedge C = 32°$. These values are slightly low for pure rhodonite but agree with those of rhodonite containing some Mg, Fe, and Ca in addition to the Mn and SiO_2 required by its formula (Larsen and Berman, 1943, p. 131). Harry Rose of the U. S. Geological Survey estimated from spectrograms that specimens from the Star and Hobo mines contain approximately 4 or 5 per cent Ca, 2 per cent Fe, and 1 or 2 per cent Mg, in addition to abundant manganese.

The textural relations in thin sections indicate that rhodonite has replaced the pyroxenes. In Figure 1 of Plate 3, which shows material from the 1950-foot level of the Star mine, needles of rhodonite occur in quartz at the end of the pyroxene, and a vein of rhodonite cuts the early pyroxene. Small quartz veins cut across the pyroxene and the rhodonite. The change of a pyroxene containing Ca, Mg, Fe, Mn, and SiO_2 to rhodonite with a composition of $MnSiO_3$ would liberate Ca, Mg, Fe, and probably some silica. Schaller (1938, p. 576) suggested that the hydrous calcium silicate xonotlite, at Pueblo, Mexico, was formed under these conditions. No xonotlite has been observed at Vanadium, New Mexico, but the calcite, which occurs at the Star mine between rhodonite and pyroxene (Pl. 3, fig. 2), may represent calcium liberated from pyroxene. At the same locality large crystals of rhodonite occupy space formerly taken by pyroxenes (Pl. 3, fig. 3). At the Hobo mine terminated crystals of rhodonite have grown at

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Plate 3.—Photomicrographs of Rhodonite

Figure 1.—Rhodonite (R) replaces pyroxene (P). Q is quartz (white). QV is a tiny quartz vein. RV is rhodonite cutting pyroxene (P). 1950-foot level, Star mine, Vanadium, New Mexico.

Figure 2.—Rhodonite (R) replaces pyroxene (P). C is calcite. 1950-foot level, Star mine, Vanadium, New Mexico.

Figure 3.—Rhodonite crystals (R) developed in pyroxene (P). 1950-foot level, Star mine, Vanadium, New Mexico.

Figure 4.—Rhodonite crystals (R) in quartz (Q) formed from pyroxene (P). Hobo mine, Vanadium, New Mexico.

Figure 5.—Rhodonite (R) replaces pyroxene (P). Hobo mine, Vanadium, New Mexico.

Figure 6.—Rhodonite (R) replaces pyroxene (P). Princess mine, Vanadium, New Mexico.

Plate 4.—Photomicrographs of Manganoan Hedenbergite, Cummingtonite, and Ore Minerals

Figure 1.—Manganoan hedenbergite (H) shows bladed structure. Used for analysis (Table 2, no. 4). Ninth level, San Antonio mine, Santa Eulalia, Chihuahua, Mexico.

Figure 2.—Cummingtonite (CU) shows fibrous structure. Used for analysis (Table 3, no. 1). Tenth level, San Antonio mine, Santa Eulalia, Chihuahua, Mexico.

Figure 3.—Manganoan hedenbergite (H) altered along cleavage to cummingtonite (CU). San Antonio mine, Santa Eulalia, Chihuahua, Mexico.

Figure 4.—Manganoan hedenbergite shows columnar structure used for analysis (Table 2, no. 5). 1400-foot level, Hidalgo mine, Santa Barbara, Mexico.

Figure 5.—Magnetite (M) and sphalerite (S) are later than pyroxene (P). 1950-foot level, Star mine, Vanadium, New Mexico.

Figure 6.—Magnetite (M) and sphalerite (S) cut and fill space in pyroxene (P) and rhodonite (R). 1950-foot level, Star mine, Vanadium, New Mexico.

the expense of pyroxene (Pl. 3, fig. 4). Rhodonite replaces columnar pyroxene in a sample from the Hobo mine (Pl. 3, fig. 5) and in another from the Princess mine (Pl. 3, fig. 6). These relations were photographed in plane-polarized light, but replacement features stand out even more clearly under crossed nicols because the birefringence of rhodonite is low compared with that of the associated pyroxenes.

The observation that ferroan johannsenite at the Star mine with 10.9 per cent MnO is replaced by rhodonite as easily as johannsenite at the Hobo mine with 25.9 MnO strongly supports the petrographic evidence that most of the MnO for the formation of rhodonite was introduced by the ore fluids rather than supplied by the alteration in place of the pyroxenes with small amounts of manganese. The source of the ore fluids was probably the intrusive magma of the Ground Hog district, which invaded the limestones of Paleozoic age and formed replacement deposits with a typical pyrometasomatic mineral assemblage.

Occurrence and Properties of Cummingtonite

The presence of amphiboles in association with pyrometasomatic deposits of Mexico and New Mexico has been noted by several geologists. Spurr, Garrey, and Fenner (1912, p. 466) reported actinolitic hornblende at the Dolores mine, Matehuala, Mexico, and Spurr and Garry (1908, p. 711) listed actinolite at Velardena, Mexico. Reber (1916, p. 544) states that in the diabase dike rocks at Clifton-Morenci, Arizona, augite is changed to hornblende or to actinolite and serpentine; that amphibole (tremolite) is found in metamorphosed limestone with andradite, epidote, diopside, magnetite, specularite, and pyrite; and that amphibole (actinolite) is found in metamorphosed shales with epidote and quartz.

Smythe (1921, p. 416) observed a dark-blue amphibole in association with hedenbergite and garnet near Fairview, New Mexico. Schmitt (1939, p. 794, 796) described ferro-tremolite as an alteration product of hornblende of igneous rocks and as an alteration of salite in the Pewabic mine. Loughlin and Koschmann (1942, p. 100) list actinolite among the minerals occurring at Magdalena, New Mexico. Hewitt (1943, p. 184) noted fibrous silicates—tremolite, actinolite, and possible hedenbergite—ranging from minute spherules to coarse nodules 2½ inches in diameter at the San Antonio mine, Chihuahua, Mexico, in association with ilvaite. Koch (1956, p. 22) considers that actinolite is widespread at the Frisco mine, Chihuahua, Mexico, but is not so conspicuous as diopside and is probably less abundant. He says that the composition of the amphibole ranges from the pure magnesium end member tremolite to actinolite with a ratio of Mg:Fe of about 4:1. Chemical analyses of these amphiboles are not available; their CaO content is therefore unknown.

Amphiboles were collected by Allen from the Kearny, Princess, Star, and Pearson mines, Empire Zinc Company's mine, Fierro dump, Linchburg Tunnel, and the San Antonio mine. Specimens from three of these localities, the San Antonio, Pearson, and Kearny mines, were carefully prepared for chemical analysis (Table 3, nos. 1, 2, 3). In all these the CaO is too low for actinolite, which ranges from 11.44 to 16.68 in the analyses of actinolite listed by Dana (1892, p. 393). This suggests that these amphiboles belong to the cummingtonite series, which is made up of a magnesium end member, $Mg_7(OH)_2Si_8O_{22}$, and an iron end member, grunerite, $Fe_7(OH)_2Si_8O_{22}$. The SiO_2 of these amphiboles is higher than that in the type cummingtonite (Table 3, no. 4) and in the cummingtonite from Baltimore, Maryland (Table 3, no. 5).

These amphiboles were formed from pyroxenes. All stages can be observed from pyroxenes containing only a few needles of amphibole to specimens of amphibole with only a few relict grains of pyroxene. The presence of 1.01 to 2.82 MnO in the analyses of these amphiboles (Table 3, nos. 1, 2, 3) adds additional evidence that they were derived from a manganese-bearing pyroxene. Photomicrographs included here show typical occurrences of cummingtonite and its relationship to the pyroxenes and the ore minerals. At the Kearny mine cummingtonite (Pl. 1, fig. 4) is shown with its typical fibrous structure and inclusions of sphalerite and magnetite, which were removed from the sample used for the chemical analyses (Table 3, no. 3). At the Empire Zinc Company's mine (Pl. 1, fig. 5), cummingtonite is replaced by

quartz, calcite, and magnetite. At the Star mine (Pl. 1, fig. 2), isolated areas of manganoan hedenbergite extinguish under crossed nicols at the same time, suggesting relics of a large grain consisting of a single crystal. Cummingtonite separates and replaces manganoan hedenbergite. Cummingtonite was replaced by magnetite, pyrite, and sphalerite (Pl. 2, fig. 2). In the Linchburg Tunnel, Magdalena, New Mexico (Pl. 2, fig. 4), a similar amphibole is replaced by garnet, calcite, sphalerite, pyrite, and quartz. This amphibole is here called cummingtonite, because its optical properties and occurrence are similar to the three amphiboles analyzed and found to be low in CaO. At the San Antonio mine, Chihuahua, Mexico, cummingtonite occurs in radiating fibers (Pl. 4, fig. 2) up to 2 inches long that were used for the chemical analysis (Table 3, no. 1). Cummingtonite was derived from manganoan hedenbergite, as is indicated by the occurrence of cummingtonite along the cleavage of the manganoan hedenbergite (Pl. 4, fig. 3).

A fibrous amphibole derived from a pyroxene is known as uralite, and the process by which pyroxenes are changed to uralite is known as uralitization. In the pyrometasomatic zinc deposits studied the uralite formed is closely related to cummingtonite. This is an unusual situation; the writers have not been able to find other published examples of calcium pyroxenes with 20–24 per cent CaO altered to amphiboles with 2.82–3.86 per cent CaO (Table 3, nos. 1, 3). The CaO content of the amphibole, no. 2 (Table 3), is believed to be intermediate in amount between that in samples 1 and 3 (Table 3), because the chemical analysis shows 1.84 per cent CO_2; if this is combined with sufficient CaO to form calcite, it would proportionally reduce the CaO in the amphibole.

Dana (1892, p. 390) credits Gustav Rose with the observation, in 1830, that pyroxene changes to an aggregate of hornblende needles. Since that time many geologists have recognized this change and are convinced that the composition of the pyroxene is reflected in that of its derivative. Uralitization is said to involve oxidation of divalent to trivalent iron, a great loss of calcium, a gain in silicon, alkalies, and at some localities a gain in magnesium. Schwartz (1939, p. 198) considers that uralitization is apparently not characteristically hydrothermal in origin and quotes from Van Hise (1904, p. 274) that it takes place during metamorphism of various types. That it occurs in connection with pyrometasomatic deposits suggests that this change takes place at temperatures above those typical of the moderate hydrothermal stage of ore deposition. Also, the extent of alteration during uralitization varies from one deposit to another, and chemical analyses are needed to confirm the nature of the alteration and the minerals formed. At the pyrometasomatic iron deposit at Iron Mountain, Missouri, Allen and Fahey (1956) found that actinolite with 10.30 per cent CaO, 13.25 per cent MgO, and 9.68 per cent FeO is formed from salite, of which the analyzed specimen contains 20.48 per cent CaO, 13.43 per cent MgO, and 8.92 per cent FeO. Assuming that the salite from which the actinolite was formed had the same composition as that found in the published analysis of the writers' specimen of salite from Iron Mountain (1952, p. 736), an increase in SiO_2, Al_2O_3, TiO_2, Fe_2O_3, FeO, MnO, and H_2O is indicated during the uralitization at Iron Mountain. This suggests that removal of CaO during uralitization has gone further toward completion in the investigated pyrometasomatic zinc deposits of New Mexico and Mexico than in the pyrometasomatic iron deposit at Iron Mountain, Missouri.

Occurrence and Properties of Garnet

The garnet at the Pewabic mine is andradite, as shown by chemical and optical data (Schmitt, 1939, p. 791–794). The optical data given for the garnet at the Magdalena district (Loughlin and Koschmann, 1942, p. 100) are those of andradite (n = 1.8885). All the specimens of garnet examined by the writers from the Combination, Princess, Pearson, and Kearny mines have indices of refraction close to 1.90± and thus are also probably the calcium-iron garnet, andradite (Skinner, 1956, p. 429).

In many pyrometasomatic deposits the garnet is zoned and birefringent. At the Princess mine (Pl. 2, fig. 3), zoned garnet is apparently replaced by manganoan hedenbergite. At the Linchburg Tunnel, Magdalena, New Mexico, zoned garnet (Pl. 2, fig. 4) is replaced by an amphibole that is here called cummingtonite, although the available specimens were not

satisfactory for a determination of their calcium content. At the Pearson mine (Pl. 2, fig. 5) euhedral, zoned garnet is replaced by cummingtonite. The determination of the temperature at which birefringent garnets become isotropic by heating has been used to indicate the temperature below which the garnet crystallized. Merwin (1915, p. 108) found that contact garnets from Alaska lost their birefringence when heated to about 800°C. Stose and Glass (1938) determined 860°C. as the temperature at which zoned, birefringent andradite from Pennsylvania became isotropic. Ingerson and Barksdale (1943) observed that in an iridescent garnet from Nevada the birefringence did not decrease until the temperature reached 1060°C. and continued almost to the melting point.

Specimens of andradite from the Pewabic mine that were heated by the writers lost their birefringence at 860°C. On this basis, crystallization of garnet at the Pewabic mine took place at a temperature below 860°C.

Origin and Sequence of Minerals

The mineral assemblage of these pyrometasomatic zinc deposits suggests that their deposition began at a high temperature from a source connected with intrusive stocks that are visible or concealed below the area of silicate deposition. At several of the deposits, garnet seems to have been the first mineral to crystal lize. This was closely followed or overlapped in time of deposition by manganoan hedenbergite Loughlin and Koschmann (1942, p. 100) consider that garnet formed later than heden bergite at the Linchburg Tunnel, Magdalena, New Mexico. At the Pewabic mine, Schmitt (1939, p. 802) lists andradite and epidote as earlier than salite. If the temperature at which andradite from the Pewabic mine loses its birefringence on heating is a reliable indicator, the early stage of silicate deposition took place at a temperature below 860°C. At all the localities where chemical data for the pyroxenes are available, the pyroxenes contain manganese, and the most consistent varieties represented are manganoan hedenbergite and manganoan magnesian hedenbergite. Locally, manganese was supplied in sufficient quantities from the invading magma to form johannsenite. At some localities a surge of manganese formed rhodonite that replaced some of the earlier ferroan johannsenite and johannsenite.

At somewhat lower temperature but during the same general epoch of silicate deposition, uralitization of the pyroxenes took place. Lindgren (1933, p. 705) states that pyroxenes occur in the inner and amphiboles in the outer contact zone of some deposits and quotes Becke that the transition point between the stability field of the two minerals is about 550°C. at 200 atmospheres. This process of uralitization is indicated by the needles of amphibole that are included in the pyroxenes at many localities and by the radiating groups of fibrous amphiboles in which the ore minerals are embedded. If the fibrous amphibole from the San Antonio mine, Chihuahua, Mexico (Table 3, no. 1), was formed from a pyroxene of the composition of the manganoan magnesian hedenbergite (Table 2, no. 4) from the same mine, the process involved a great loss of CaO and a gain in SiO_2, FeO, Al_2O_3, and MgO. The loss of CaO from 20.46 per cent in the manganoan hedenbergite to 2.82 per cent CaO in the amphibole resulted in the formation of a variety of uralite more closely related to cummingtonite than to actinolite. The analyses of the fibrous amphiboles from the Pearson mine and the Kearny mine (Table 3, nos. 2, 3) show a similar low CaO content but have a higher MgO than FeO content and suggest the presence of relatively higher proportions of the magnesium end member of the cummingtonite series than that in the specimen from the San Antonio mine.

The sulfide ore minerals were deposited at a lower temperature and later than the silicate minerals, including the cummingtonite. In Plates 1, 2, and 4, sphalerite, magnetite, and pyrite are shown cutting and replacing manganoan hedenbergite, garnet, and cummingtonite. Schmitt (1939, p. 805) stressed the early age of the high-temperature silicates and the late age of the sulfides at the Pewabic mine. Loughlin and Koschmann (1942, p. 106) recognized that deposition of the sulfide minerals in the Magdalena district took place at moderate temperatures, compared with the higher part of the moderate-temperature range of the replacement deposits in the Kelly Limestone in which hedenbergite and garnet are found. Koch (1956, p. 36) remarks that the sulfide

minerals sphalerite, galena, chalcopyrite, pyrite, and arsenopyrite are later than the silicate minerals at the Frisco mine, Chihuahua, Mexico.

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Institute of Technology, St. Louis University, St. Louis, Mo.; U. S. Geological Survey, Washington, D. C.

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GEOPHYSICAL INVESTIGATIONS IN THE EASTERN CARIBBEAN: TRINIDAD SHELF, TOBAGO TROUGH, BARBADOS RIDGE, ATLANTIC OCEAN

BY J. I. EWING, C. B. OFFICER, H. R. JOHNSON, AND R. S. EDWARDS

ABSTRACT

A number of seismic-refraction stations were occupied in the eastern Caribbean area, and the results are correlated with other geophysical measurements and with land geology. The purpose of the investigation was to measure the crustal and subcrustal structure in the Caribbean, along the island arc, and in the Atlantic to determine the structural relationship of the interior-basin—island-arc—deep-sea-trench sequence.

The measurements indicate that the Caribbean was formerly an oceanic area which in a large part has been altered to produce a thickened crust and a lowered seismic velocity in the subcrustal material. Along the island arc the crust has been considerably thickened, and its upper surface approximately parallels the topography. On the Atlantic side of the islands the crustal and subcrustal layers are downwarped or downthrust to parallel approximately the topography across the Puerto Rico trench and its extension.

CONTENTS

TEXT

	Page
Introduction	897
Acknowledgments	898
Seismic-refraction results	898
Section A-A′	898
Section B-B′	903
Section C-C′	904
Section D-D′	904
Caribbean—Lesser Antilles—Atlantic section	905
Results not included in structure sections	906
Discussion of results	906
Island arc	906
Puerto Rico trench and Barbados ridge	907
Grenada Trough, Aves Swell, Venezuelan Basin	908
The 3.2–4.5 km/sec. layer	908
Possible southern extension of the island-arc structure	909
Origin and development of the Eastern Caribbean	909
References cited	911

ILLUSTRATIONS

Figure	Page
1. Chart showing seismic-station locations and traverses along which structure sections were drawn	899
2. Structure section from Grenada eastward to deep water in the Atlantic (section A-A′)	902
3. Structure section from Trinidad to Grenada (section B-B′)	903
4. North-south structure section along the Barbados ridge (section C-C′)	904
5. Structure section from St. Lucia to the Barbados ridge (section D-D′)	904
6. Generalized structure section from the Aves swell to deep water east of the Barbados ridge	905
Plate	**Facing page**
1. Travel-time graphs	912

TABLE

Table	Page
1. Layer velocities and thicknesses	900

INTRODUCTION

During the winter and spring of 1955, 47 seismic-refraction stations were occupied in the eastern Caribbean area to study the crustal structure in and around the Antillean orogenic belt. Measurements were made in the Puerto Rico trench, in the Venezuelan basin of the Caribbean, along the island arc, on the Barbados ridge, and on the Trinidad shelf. Previous refraction work has been done by Worzel and Ewing (1948) in the Barbados-Trinidad area, by Tolstoy *et al.* (1953) near Martinique, by Hersey *et al.* (1952) in the Puerto Rico trench, by Sutton and Ewing (In preparation) in the Venezuelan basin, around Puerto Rico, and in

the Puerto Rico trench, and by J. Ewing (Unpublished) around Barbados. This paper is primarily concerned with the stations in the southeastern part of the area; a companion paper by C. B. Officer *et al.* (1957) deals primarily with the northern part. Some overlapping of data and discussion occurs in the two papers.

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All four authors took part in the expedition and the preliminary analysis of the data. The final analysis, interpretations, and conclusions presented here are those of the first two authors.

Seismic-Refraction Results

Measurements and Interpretation

The results reported here were obtained using the standard methods of measurement and interpretation described by Ewing *et al.* (1937; 1939; 1950), Officer *et al.* (1952), Officer and Wuenschel (1951), and Sutton and Bentley (1953). The station locations and section lines are shown in Figure 1. Reversed stations are indicated by circles or crosses at both ends of the line showing the shooting track; other stations shown are unreversed. The bathymetry, showing the 500-fathom contour intervals, was taken from the U.S. Navy Hydrographic Office Chart No. 5487.

The travel-time graphs of stations 11–28 from this cruise and of stations G-5, G-6, G-7, and G-8 from J. Ewing (Unpublished) are shown in Plate I. The graphs for stations 29, 30, 31, and others discussed in the companion paper by Officer *et al.* (1957) will be published in a later report. Most of the seismic results discussed here are presented in the form of structure sections (Figs. 2–6). and all results are listed in Table 1. Section A - A′ is made up from the stations lying near a line running east from Grenada. Section B - B′ runs north from Trinidad to Grenada, C - C′ runs north from station 22 past the western side of Barbados, and D - D′ runs southeast from St. Lucia. The last is a generalized section from the Aves Swell to deep water east of Barbados.

Section A - A′

From Grenada out to about 2700 fathoms water depth east of Barbados (Fig. 2), seismic velocities in km/sec. are shown at the points of measurement. Throughout most of the section two low-velocity sediment layers are found. The average velocities are about 1.7 km/sec. for the upper layer (stippling) and about 2.4 km/sec. for the second layer (narrow crosshatching). Underneath these, along most of the section, is a layer whose average velocity is about 4 km/sec., shown by the wide crosshatching; from its velocity it could be identified as lithified sediments, volcanic rocks, or possibly metamorphosed sediments. The next higher

velocity layer is found only in the middle portion of the section. Its average velocity is about 5.1 km/sec. and could indicate intrusive oceanic stations. The seaward end of the section shows a thickness of 5–6 km, which is normal for oceanic crust, and although the layer cannot

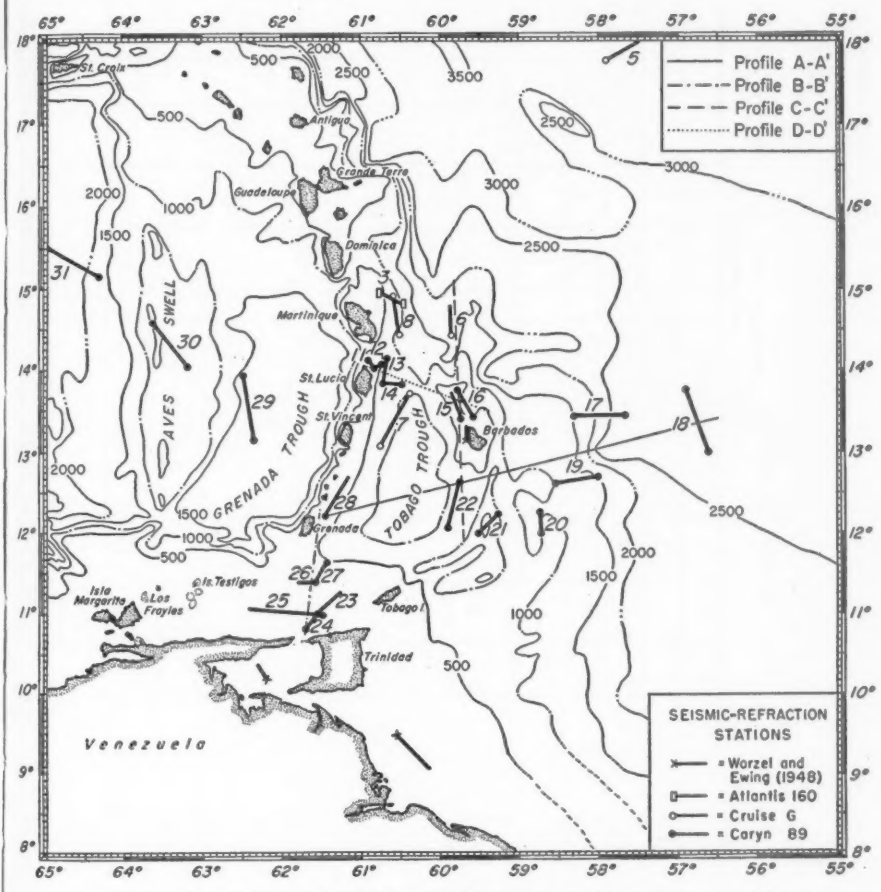


FIGURE 1.—CHART SHOWING SEISMIC-STATION LOCATIONS AND TRAVERSES ALONG WHICH STRUCTURE SECTIONS WERE DRAWN

or metamorphic rocks. The layer is not found at either end of this section, and apparently does not extend under the island arc here. At other places around the arc a similar velocity is found under or near the islands.

The next deeper layer has an average velocity of 6.7 km/sec. and is identified as the high-velocity crustal layer commonly found at be followed across the middle portion of the section, it appears again at the Grenada end considerably thickened. The measurement near Grenada is unreversed but was made with the shot line running parallel to the bottom contours and should give a reasonably accurate velocity. The second station from the left (G-7) is reversed and shows almost the same velocity

TABLE 1.—LAYER VELOCITIES AND THICKNESSES
(In km and km/sec.)

		Layer Velocity								Layer Thickness						
Station	Ocean	A	B	C	D	E	F	G		Ocean	A	B	C	D	E	F
G- 5	1.51	(1.73)					6.82	8.23	E. W.	5.49 5.49	0.74 0.76					5.92 7.44
17	1.50	(1.70)	2.58				6.74	8.32	E. W.	4.07 2.85	1.20 0.54	2.62 6.21				5.79 6.79
18	1.51	(1.70)	2.23				6.64	8.09	N. S.	5.05 4.73	0.25 0.11	1.99 2.99				7.07 3.62
19	1.50	(1.70) + 1.91	2.41		4.31		6.77*		E. W.	2.37 2.01	2.38 0.67	2.87 5.37		3.10* 3.92*		
G- 6	1.51	(1.74)	2.73*	3.35*			6.17 to 6.88*		N. S.	3.14*	0.86*	2.67*	3.80* to 5.09*			
15	1.50	1.81*	2.50*	3.38*					N. S.	0.73*	0.78*	2.18*				
16	1.50	(1.72)	2.22	3.21	4.27				N. S.	0.37 1.09	0.38 0.27	2.00 0.51	5.20 3.32			
Barbados†	1.50	1.71*	2.75*	3.38*					N. S.	0.05*	0.75*	0.95*				
20	1.50	(1.70)	2.95*		[3.70]				N. S.	1.27 2.10	[5.63] 0.87	[2.23] 2.23*				
21	1.50	(1.70) + 1.93			3.81	5.31			N. S.	2.19 2.19	3.48 2.17			4.70 3.39		
22	1.50	(1.70)			3.97	4.91	(6.74)		N. S.	1.10 1.23	1.07 0.32			2.56 2.82		
Orinoco (Shoal)†	1.50	1.76*	2.43*			5.15*			N. S.	0.02*	0.55*	3.12*				
Orinoco (Deep)†	1.50	1.67*	2.00* + 2.37*	3.43*			6.62*		N. S.	0.04*	0.48*	1.53*	4.85*			
		(1.66)							N.							

23	1.52	(1.66) + 2.01*	2.71*	3.68*		5.10*	6.83*		N. S.	0.19	1.03	2.51	0.34		2.19	
24	1.52	(1.66) + 1.93	2.75	3.43		4.80	6.38		N. S.	0.19	1.03	2.51	0.34		2.19	
25	1.52*	(1.70) + 2.01	2.93*			5.20*	6.47*		S.	0.19	0.81	0.37	1.74		3.03	
									E. W.	0.26*	1.24*	2.95*			2.18*	
26	1.51	1.64* 2.03*	2.57*		4.33*				E. W.	0.18	1.29	2.35				
27	1.51	1.64 + 2.03	2.57		4.33				N. S.	0.68 0.18	1.66 1.29	1.93 2.35				
A-160-3††	1.5	(1.7)	2.6	3.9							?	?	?			
G- 7	1.50	(1.72)	2.86		4.01		6.84		N. S.	1.72 2.09	1.24 0.95	1.57 1.98		4.80 2.72		
8	1.50	1.68	2.64		4.07		6.62		N. S.	2.12 1.58	1.06 0.27	2.12 1.70		3.12 1.96		
11	1.54	1.76	2.68	3.44	4.40		6.31		N. S.	0.08 0.08	0.05 0.13	0.96 0.77	0.41 0.88	0.62 2.49		
12	1.52	1.71	2.39	3.25	4.29		5.97		E. W.	0.95 0.09	0.61 0.11	0.18 0.67	1.25 0.85	1.51 2.56		
13	1.50	(1.70)			4.33	5.30			N. S.	1.10 0.82	1.30 1.30			3.48 2.96		
14	1.50	(1.70)			4.19	5.35			E. W.	1.54 0.82	1.48 1.31			3.20 2.63		
28	1.51	(1.70)	2.33*		4.00*		6.40*		N. S.	0.04*	0.06*	1.34*		3.05*		
29	1.51	2.00		3.82		5.29	6.34	7.47	N. S.	2.93 2.93	0.66 1.37		2.05 2.88		1.94 2.69	5.64 8.70
30	1.51	(1.80)		3.76	4.33	6.02	6.63		N. S.	1.68 1.30	0.36 0.60		1.49 1.45	3.89 2.78	1.50 4.07	
31	1.51	1.86	2.81*		4.47	5.96*	6.58	7.36	N. S.	4.28 3.44	1.72 0.86	1.08*		1.57 3.22	3.05*	2.79 7.83

A—Unconsolidated sediment: velocity range 1.6–2.0 km/sec.
B—Partially lithified to lithified sediment: velocity range 2.0–3.0 km/sec.
C—Lithified sediments, volcanic rocks, metamorphic rocks: velocity range 3.2–3.8 km/sec.
D—Volcanic or metamorphic
E—Metamorphic or igneous
F—Crustal velocities
G—Velocity beneath major discontinuity
N., S., E., W.—Locations of receiving positions relative to each other
†—From Worzel and Ewing (1948)
††—From Tolstoy *et al.* (1953)

Velocities with asterisks are unreversed. Thicknesses with asterisks have been computed with unreversed measurements. Assumed velocities are enclosed in parentheses and in most cases are governed by measured values near by. Numbers in brackets indicate uncertain velocities or thicknesses obtained from a complex interpretation.

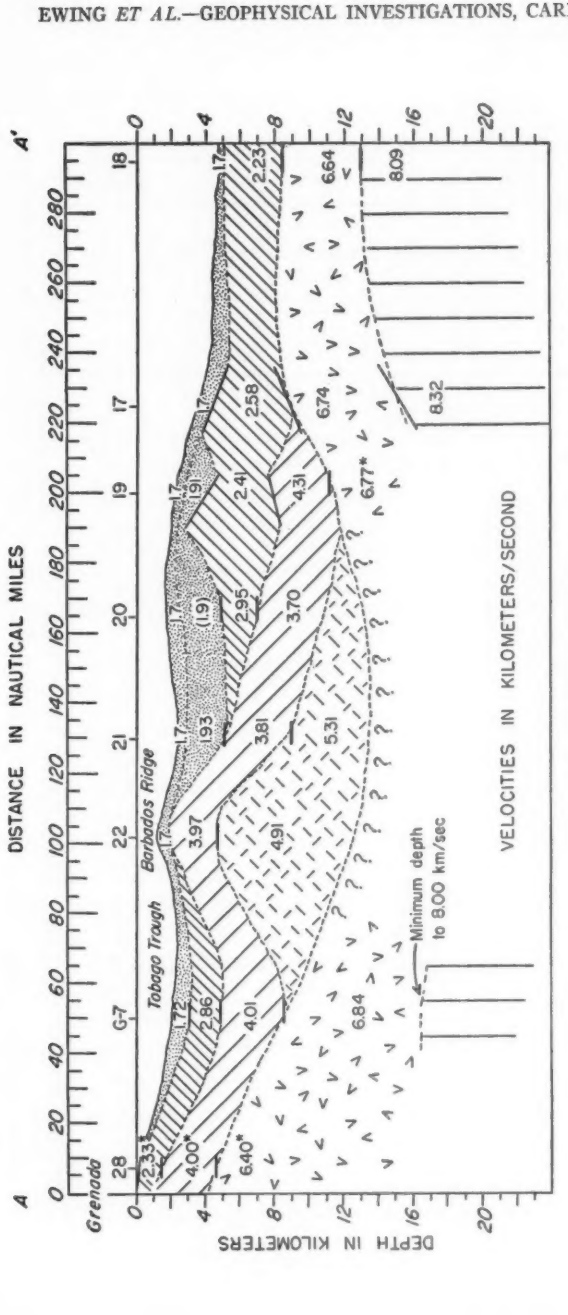


FIGURE 2.—STRUCTURE SECTION FROM GRENADA EASTWARD TO DEEP WATER IN THE ATLANTIC
(SECTION A-A')
* Unreversed velocity

as at the seaward end. Thus oceanic crust is found at both ends of this section dipping down toward the center; it was not found in the central part, probably because it is too deep there to be detected without larger shots and longer profiles.

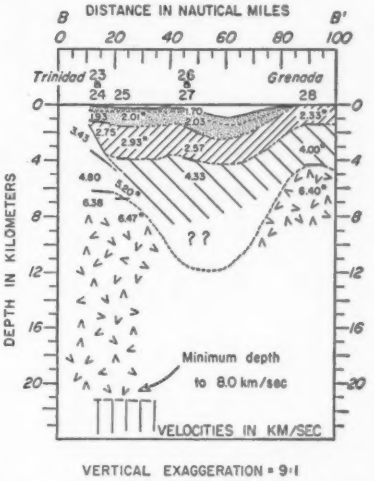
The mantle was reached at the two seaward stations only. At station 18 it is found at about 13 km depth and at station 17 it is approximately 2 km deeper and has a 1.5° westward dip.

The structure in the middle portion of the section is probably oversimplified. Most of the stations around the Barbados ridge are complicated; they show fairly large dips and in some cases horizontal discontinuities which indicate severe folding and faulting. A striking example is station 20, where the G_3 line is offset by about 2.5 sec. near the north end. It is not possible to give more than an approximate structural interpretation here. Travel times indicate that the highest-velocity layer slopes upward from south to north and has been faulted or folded down (north relative to south) by several kilometers in the region of the offset. Because of complications like this, the results given for some Barbados ridge stations may have errors in velocity determinations as high as 15 per cent and even larger errors in depth determinations. The reader can estimate the possible error at individual stations by examining the travel-time graphs (Pl. 1) for complexity and scatter of points.

This section shows the high-velocity crustal layer and the subcrustal material to be downfolded or down dropped in the Barbados ridge area and overlain by a considerable amount of low-velocity sediments and a large thickness of material in which the velocity of sound is 4–5 km/sec. The upwarping of the deeper layers under Grenada shows in cross section a raised crustal platform under the island arc.

Section B - B′

Figure 3 shows a north-south section between Trinidad and Grenada. In the middle of the section a large syncline is indicated but not positively determined. Station 27, located about midway between B and B′, was heavily shot, but no velocity higher than 4.3 km/sec. was determined by first arrivals, even by the longest shots which were at a range of about 30 miles. Second arrivals give some indication of a 6.4 km/sec. layer at a depth of about 12 km, but the evidence is not good enough to warrant drawing lines on the travel-time graph. However, the failure of the travel-time graph to show a break from the 4.3 km/sec. layer means that the depth to a high-velocity basement must be on the order of that shown.



VERTICAL EXAGGERATION = 9·1

FIGURE 3.—STRUCTURE SECTION FROM TRINIDAD TO GRENADA (SECTION B-B′)
* Unreversed velocity

To the right (north) in this section the higher-velocity layers rise under the Lesser Antilles arc. To the left (south) they rise similarly under northern Trinidad. The synclinal or downdropped area between is possibly a southwestward-trending branch of the main downbuckle under the Barbados ridge. Near Trinidad the structure becomes more complex. The preferred interpretation of the travel-time data of stations 23, 24, and 25 is shown in Figure 3. This shows the 2.3–2.9 km/sec. sedimentary layer almost pinching out in station 24, as both the 3.4 km/sec. layer and 4.8–5.2 km/sec. layers slope up rather sharply southward. These upward-sloping layers probably represent the northern flank of the Northern Coast Range of Trinidad and Venezuela.

The alternate interpretation of the data results in approximately the same structure,

except that the 3.4 km/sec. layer and the 2.7 km/sec. layer would be included in a single layer with a velocity of about 2.9 km/sec. In either case the 6.4 km/sec. crustal layer is at about the depth shown.

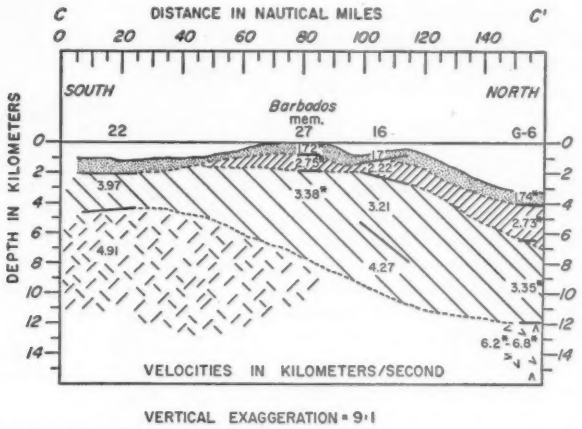


FIGURE 4.—NORTH-SOUTH STRUCTURE SECTION ALONG THE BARBADOS RIDGE (SECTION C-C')
* Unreversed velocity

Station 25 was long and heavily shot, and minimum depth to the mantle was determined as 21 km, a reasonable figure for a station near the edge of a continent.

Section C - C'

Figure 4 shows a north-south section passing just off the west shore of Barbados. The high-velocity crustal layer was not reached at any of the stations on the shallow part of the ridge. The northernmost station, G-6, shows it to be at a depth of 10–12 km. Possibly it is even deeper, since the shots at the station were fairly widely spaced at the longer ranges because of misfires, and a thickness of up to 2 km of the 4.5 to 4.9 km/sec. layer could have been missed between the 3.35 km/sec. layer and the crust. Although the velocity in the deep crustal layer is not well determined, the two lines, $G_4^{(a)}$ and $G_4^{(b)}$, on the travel-time graph show reasonable limits. Interpretation of the data at the Barbados ridge stations is not straightforward, as can be seen in the travel-time graphs. Complexity of travel-time data indicates complexity of layering.

Section D - D'

Figure 5 shows a section running southeast from St. Lucia along the ridge which bounds the Tobago trough on the north. Correlation

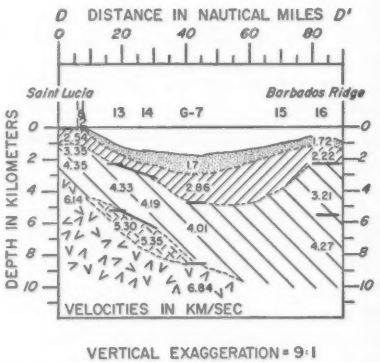


FIGURE 5.—STRUCTURE SECTION FROM ST. LUCIA TO THE BARBADOS RIDGE (SECTION D-D')

of the deeper layers along this traverse is uncertain, and the interpretation shown may be incorrect in detail. As in section A - A', the highest velocity measured was the 6.0 – 6.3 km/sec. at stations 11 and 12 just offshore from the island arc. Although the shot lines at these stations were relatively short and therefore may not give an accurate determination of the

velocity in the deepest layer, the measured velocity probably corresponds to the 6.5 km/sec. oceanic crustal layer, perhaps somewhat altered to account for the subnormal velocity. There is no control on the interface between the 5.3 km/sec. and the 6.1 km/sec. layers east of station 12, but there are indications that the 5.3 km/sec. layer is relatively thin under stations 13 and 14.

The crust and subcrust under the Barbados ridge are believed to be deformed into a large northward-trending syncline. This was indicated first by section A - A′ and further by the other stations on the ridge that fail to show a velocity higher than about 5 km/sec. for the deepest layer detected.

Caribbean-Lesser Antilles-Atlantic Section

Figure 6 shows a generalized section running from the Aves swell in the Caribbean across the Lesser Antilles and the Barbados ridge and out to deep water in the Atlantic. This does not represent a particular traverse and is primarily intended to show the general structural relationship of the island arc to the oceanic and the Caribbean areas. Details of the structure under the islands are not known but may be more complicated than this drawing shows. The same is probably true of the Barbados ridge and some parts of the Aves swell.

The upper two layers, one the unconsolidated sediment with velocities around 1.7 km/sec. and the other the partially lithified sediments with velocities from 2.0 to 2.9 km/sec., have been combined into the stippled layer. Sediments from the Orinoco river transported by turbidity currents may have contributed appreciably to the accumulation of the great thickness of low-velocity deposits in the Tobago trough and particularly on the deep-sea floor east of the Barbados ridge.

The 3.5–4.5 km/sec. layer, shown by cross-hatching, is probably less uniform than the drawing indicates; it is probably not a continuous geologic sequence across the area, but the presence throughout the area of material with this velocity range is significant. These velocities are characteristic of a wide variety of materials from lithified sediments and volcanic rocks to some metamorphic rocks. Along the Lesser Antilles and the Aves swell this layer is

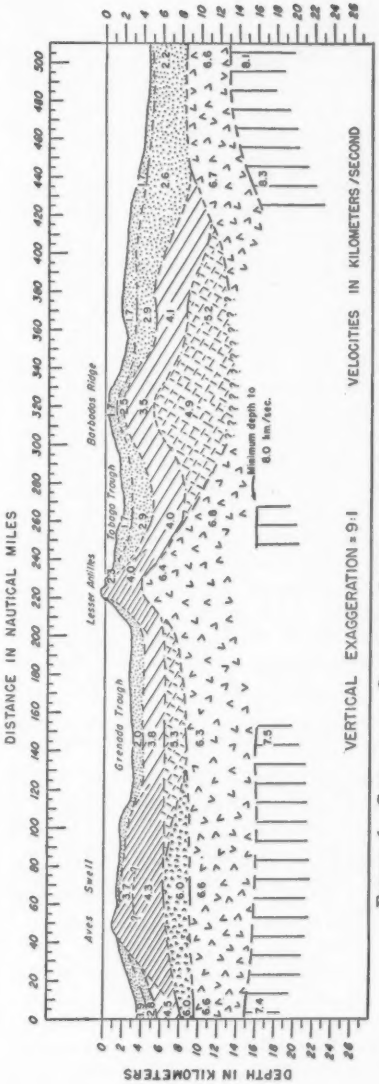


FIGURE 6.—GENERALIZED STRUCTURE SECTION FROM THE AVES SWELL TO DEEP WATER EAST OF THE BARBADOS RIDGE

probably primarily composed of extrusive volcanic material; under the Grenada trough, Tobago trough, and Barbados ridge it is more likely to be a combination of lithified pyroclastics and other sediments. The 4.9–5.3 km/sec. material is found only in the Grenada trough and under the Barbados ridge in this section, but at other locations around the arc a similar velocity is found on and near the islands. This velocity range is characteristic of metamorphic or intrusive igneous rocks.

The high-velocity crustal layer was traced across the entire area, with the exception of the Barbados ridge. Although a tendency is noticed toward a lower velocity and some differentiation or contamination in the Caribbean, the widespread occurrence of this layer, with a velocity characteristic of oceanic crust, indicates that the Lesser Antilles island-arc structure resulted from tectonic forces in a formerly oceanic, rather than continental, area. Ewing and Worzel (1954) and Ewing and Press (1955), reporting the preliminary results from earlier seismic work in the Caribbean, point out that the crustal structure is of oceanic rather than continental type. Subcrustal material on the Atlantic side of the arc has a velocity normally associated with the mantle, but an appreciably lower velocity is found for the subcrust[1] under the Caribbean. This contrast and its structural implications are discussed below and also by Officer *et al.* (1957).

Results Not Included in Structure Sections

Although stations G-5 and G-8, station A-160-3, and the two Orinoco delta stations of Worzel and Ewing (1948) were not included in the preceding structure cross sections, the results (Table 1) are not contradictory to those included.

Discussion of Results

Island Arc

From Puerto Rico to Grenada, and possibly to Margarita, the islands of the eastern Caribbean lie along an extensive arcuate belt of positive free-air gravity anomalies (Hess, 1938). Prominent gaps in the island chain are the Anegada passage, which separates the Greater Antilles from St. Croix and the Lesser Antilles, and the Dominica passage. The Lesser Antilles in the north have been divided into two groups: the outer or limestone Caribbees and the inner volcanic islands. South of Dominica there is only the volcanic group of islands. The Greater Antilles are older and more severely folded, metamorphosed, and intruded than the Lesser Antilles. The oldest dated rocks of the Greater Antilles appear to be progressively older toward the west—Jurassic in Cuba and Cretaceous or possibly Jurassic in Hispaniola and Puerto Rico (Schuchert, 1935, p. 455, 471, 495). In the Lesser Antilles the limestone Caribbees are believed to be older than the inner group. On St. Bartholomew and St. Martins tuffs and breccias of Eocene age are reported (Christman, 1953); Oligocene on Martinique (Senn, 1940) and Oligocene or early Miocene or Carriacou (Trechman, 1935) are the oldest reported for the inner volcanic islands. Volcanic activity apparently ceased in the outer group in Oligocene and was followed by erosion, submersion, and deposition of the Oligocene-Miocene limestones, whereas activity has continued in most of the inner group up to the present.

The volcanic and intrusive rocks throughout the Greater and Lesser Antilles are predominantly andesites and diorites, although basalts and ultramafic rocks are found in varying amounts. All the islands of the arc show strong short-period magnetic anomalies (Davidson and Miller, unpublished) in sharp contrast to the relatively smooth field in the surrounding area. High anomalies are also associated with the Dutch Leeward Islands, the Northern Coast Range of Trinidad, Tobago, and the Aves swell. Measurements show that no anomaly is associated with the north wall of the Puerto Rico Trench or its extension to the southeast, and none is associated with Barbados (E. T. Miller and M. Ewing, in preparation).

The seismic results (Fig. 2, 5, 6) show that the topographic ridge, along which the southern end of the Antillean island arc is located, is the surface expression of a prominent ridge in the underlying crust (6.3–6.8 km/sec. velocity). Associated with the elevation of the crust is an apparent decrease in the seismic velocity.

[1] As used throughout this paper, subcrust refers to the material directly beneath the 6.5 km/sec. layer.

The writers believe that the uplift and the associated volcanism come primarily from crustal thickening resulting from extensive intrusion from below by material with crustal velocity and density. The islands are believed to be composed dominantly of the volcanic rocks and the lighter, more siliceous components of the intruding material.

Puerto Rico Trench and Barbados Ridge

Previous seismic work in and around the Puerto Rico trench consisted of three stations by Hersey *et al.* (1952), and seven stations interpreted by G. H. Sutton which have been reported in brief (Ewing and Worzel, 1954). The results of the early work, as well as of recent work (Officer *et al.*, 1957), can be summarized as follows: The structure just north of the trench (½–1 km of sediment and 3 km of 6.6 km/sec. oceanic crust overlying the M discontinuity at a depth of 9–10 km below sea level) is approximately standard oceanic structure. It deviates from the standard in that the crust is only a little more than half normal thickness, and the M discontinuity is higher by about 2 or 3 km. In the trench the crust is 2–3 km thicker than normal and is underlain by the M discontinuity at a depth of 18 km. Above the crust is a thickness of material whose velocity is characteristic of lithified sediments, volcanic rocks, or metamorphic rocks. Between the trench and the slope to the islands the highest velocity measured was around 5.5 km/sec.

Southeastward around the Lesser Antilles, the Puerto Rico trench gradually decreases in depth and disappears. In its place the northern extension of the Barbados ridge begins to emerge and continues to the south as a prominent topographic feature. According to the bathymetry in the area, the axis of the Barbados ridge is not a direct continuation of that of the Puerto Rico trench extended but lies between the axis of the trench and the Lesser Antilles. In this area two stations have been occupied before those reported here: one near the west coast of Barbados (Worzel and Ewing, 1948) and the other in 1000 fathoms northeast of Martinique (Tolstoy, Edwards, and Ewing, 1953). Although neither of these stations was long enough to detect deep layering, the results are in good agreement with more recent work. Results reported here show the main features of this part of the orogenic belt. In the Barbados ridge area, which is near the axis of the negative gravity belt, the structure consists primarily of a great thickness of severely folded and faulted rocks with velocities between 3.2 and 5.2 km/sec.; this is believed to be a sequence of lithified sediments and volcanic rocks, possibly metamorphosed and probably intruded in the deeper parts. In the Tobago trough and east of the Barbados ridge the higher-velocity materials are appreciably deeper than along the crest of the ridge and are overlain by 3–5 km of unconsolidated and lithified sediments. Oceanic crustal velocities were found near Grenada and St. Lucia and offshore about 100 miles east of the ridge. Near the islands stations 28 and G-7 show this layer dipping east, and the offshore stations 17 and 18 show it dipping west. These results, coupled with the inability to detect this layer under the Barbados ridge, are evidence that the crust and presumably the upper mantle are deformed into a large syncline 200 miles wide with its axis centered about on the Barbados ridge (Figs. 2, 6). The structure is not known in detail under the Barbados ridge or under the Lesser Antilles, but the gross features must be approximately as shown.

Although the Barbados ridge and the Lesser Antilles are similar topographically, they are different in composition. The Lesser Antilles are formed by a rise in the underlying high-velocity crustal material; the Barbados ridge is formed by a rise in the overlying lithified sediments, volcanic and intrusive rocks, which lie within a broad crustal syncline. The Lesser Antilles are actively volcanic and highly magnetic; the Barbados ridge is not.

Complementary to the results discussed above is the observation from earthquake studies that, although no deep-focus shocks and only four intermediate ones are plotted, the seismicity of the Antillean area appears to conform to the type of the Pacific arcs (Gutenberg and Richter, 1954, p. 38). The epicenters of most shallow quakes are within the negative gravity zone, whereas three of the deeper earthquake epicenters are along the island arc, and the fourth is slightly inside the axis of the negative gravity zone.

Grenada Trough, Aves Swell, Venezuelan Basin

One station in the Grenada trough, one on the Aves swell, and one in the Venezuelan basin are shown in the section of Figure 6. Other Venezuelan basin stations are described by Officer *et al.* (1957) and by G. H. Sutton and M. Ewing (In preparation).

The Aves swell is a volcanic ridge and owes its elevation to a thickening of the 3.7–4.3 km/sec. material and to a rise and probable thickening of the crust. The crust is thinner in the Grenada trough and still thinner in the Venezuelan basin, which probably accounts for the deeper water in these areas. The layer above the crust in the Grenada trough has a velocity of 5.3 km/sec. much like the deep material under the Barbados ridge. No velocity higher than 6.6 km/sec. was found at the Aves swell station, and the subcrustal velocities found at the others are 7.4 km/sec. in the Venezuelan basin and 7.5 km/sec. in the Grenada trough. These velocities are appreciably lower than the velocity usually associated with the upper mantle that is found at approximately this depth in the ocean basins. Thus the Caribbean, at least part of it, is not typically oceanic in structure. This should not be interpreted to mean that the Caribbean structure is continental. It is approximately like the oceanic structure in crustal velocity and thickness but differs in the velocity below the deep discontinuity.

Other areas where measurements have shown a velocity similar to that found for the subcrust in the Caribbean are: along the continental margin of eastern North America (Officer and Ewing, 1954; Bentley and Worzel, 1956; J. Ewing and M. Ewing, unpublished; the Tonga ridge (Raitt *et al.*, 1955); Bermuda rise (Officer *et al.*, 1952); Mid-Atlantic Ridge and Norwegian Sea (J. Ewing and M. Ewing, unpublished); northeast of Eleuthera in the Bahamas (Katz and Ewing, 1956). In these areas, with the possible exception of the Tonga ridge, the normal oceanic crustal layer (6.5 km/sec.) is absent. In most places the 7.1–7.7 km/sec. material is overlain by a 5.0–5.5 km/sec. layer, except on the deep parts of the Mid-Atlantic ridge, where the overburden appears to consist largely of unconsolidated sediments. In the Venezuelan basin a section with normal oceanic crustal velocity (6.0–6.6 km/sec.) overlies the 7.2–7.7 km/sec. material.

The 3.2–4.5 km/sec. Layer

Throughout most of the area covered by these measurements a layer was found in which the velocity determinations ranged from 3.2 to 4.5 km/sec. (the layer shown by the wide cross-hatching). Identification of the layer from velocity considerations only is difficult, since the layer falls within the range of many lithified sediments, volcanic rocks, and some metamorphic rocks. Figure 6 shows it pinching out east of Barbados, but in deep water, a layer 1–2 km thick with this velocity would be masked by the water-wave arrivals and might not be seen as second arrivals. Recent sub-bottom reflection data and data on transformed shear waves from several deep-water stations have indicated that such a layer, lying just above the crustal layer, may be present in many parts of the Atlantic Ocean (J. Ewing and M. Ewing, unpublished.[2]) The presence of the layer in many parts of the Caribbean but not throughout the Atlantic might be accounted for by the fact that where the water is shallower it is not masked.

The writers believe that the 3.2–4.5 km/sec. layer is the result of widespread volcanism and intrusion. Since the layer apparently has no correlation with magnetic anomalies (M. Davidson and E. T. Miller, unpublished), it probably is not a massive flow but a mixture of lithified volcanic ash, tuff, breccia, sediments, and possibly thin flows. A layer with approximately this velocity is found underneath the low-velocity sediments near Bermuda (Officer *et al.*, 1952) and has been accounted for in the same way.

[2] In these measurements the horizon from which the strongest sub-bottom reflections are received and the horizon at which the transformation from compression to shear occurs are consistently shallower than the top of the crustal layer, as determined by the refracted compressional arrivals. Thus, either there is an additional intermediate layer between the low-velocity sediments and the crust, which is not seen as a refractor, or the upper 1–2 km of the crust has a velocity gradient.

Possible Southern Extension of the Island-Arc Structure

The seismic section B-B′ indicates that the Lesser Antilles volcanic ridge does not continue southward from Grenada to Trinidad. From this and from topographic considerations, it therefore appears that the Lesser Antillean ridge extends south and west from Grenada through Testigos and Los Frayles and probably on to Margarita. From observations of southward thrusting and overfolding in the Venezuelan Coast Range and in the metamorphic rocks south of Caracas and observations of northward overturning on Margarita and along the coast north of Caracas, Maxwell and Dengo (1951) postulate a northeastward-trending orogen whose axis passes south of Caracas, through the Cariaco trough, and between Margarita and the mainland. Thus this is excellent agreement of conclusions derived from field geology and from geophysical methods.

There are insufficient measurements to determine what relationship the broad synclinal development east of the Lesser Antilles has to the Venezuelan geosyncline. Possibly this broad feature, on approaching South America, is split into two smaller branches; one passes between Grenada and Trinidad, and the other bends around Tobago and joins the Venezuelan geosyncline. If this synclinal development and the negative gravity belt associated with it continue into South America, a comparison may be drawn between this and the apparent continuation of the negative strip of the East Indies into Burma (Evans and Crompton, 1946).

Origin and Development of the Eastern Caribbean

The remainder of this paper is speculation on the physical and chemical processes which might have produced the structural features described above. For other theories on the origin of the Caribbean structure, the reader is referred to Eardley (1954).

In the development of the Caribbean area from oceanic structure to the existing structure, volcanism and intrusion have obviously been of major importance. This is evident from widespread occurrence of volcanic and intrusive rocks throughout the exposed land portions of the area and from the presence in most places of seismic layers whose velocities are characteristic of volcanic and intrusive material. Although the volcanism and intrusion may be secondary processes in the tectonic development of the area, the writers believe that they are the primary processes and that they occur on such a scale that a large area of the earth's crust may be altered in a manner that results in the creation of island-arc structure. The seismic results show that large areas of the Caribbean crust are thickened and have lower than normal velocity in the upper portion; the velocity in the crust is appreciably lower than is normal for the upper mantle. These features of the seismic results strongly suggest that the crust and upper mantle have been contaminated by the introduction of a lower-velocity material. This probably occurred from widespread intrusion by a primary differentiate which migrated upward from deep in the mantle in a manner like that referred to by Poldervaart (1955, p. 138) as degassing of the mantle.

A most significant feature of the results is the presence of material with crustal velocities, 6.1 to 6.4 km/sec., high in the seismic sections under the Lesser Antilles. Since these islands are mainly built of material of andesitic composition, the seismic sections (Figs. 2, 6) should provide clues to the processes involved in the formation of such material. One possibility is that the andesitic material is the result of fractional crystallization from an olivine basalt magma. According to Turner and Verhoogen (1951, p. 126) the possibility of such an origin is generally accepted. This might be a plausible process here, since the crustal material, presumed to be olivine basalt, underlies the Lesser Antilles ridge and is at an elevated position there. However, the source of heat necessary to raise the temperature in the layer sufficiently to allow differentiation is not in evidence. Furthermore, although laboratory experiments have shown that it is possible to derive the full range of igneous rocks from a basalt melt, field evidence (Tilley, 1950; Waters, 1955) indicates that where thick sills of basalt have differentiated, they form feldspathic gabbros and end-stage granophyres but not intermediate andes-

ites. In addition, it is difficult to imagine how the total volume of extrusive and intrusive material along the Lesser Antilles ridge and the Aves swell and in the Venezuelan basin could have been derived solely from the thin 5 km oceanic crust, particularly since observations show that where the largest amounts of andesitic material are found, the remaining crust is thicker than normal.

A second possible process of deriving andesites might be by contamination of an upward-migrating basalt magma as it passed through the sediments. Andesites may have been produced in this way here early in the volcanic cycle when perhaps the entire area was overlain by a thick sedimentary cover, but recent volcanism is occurring in the zone of thinnest sedimentary cover, where contamination would be at a minimum.

A third possible process, the one preferred here, involves differentiation and separation of the lighter components deep in the mantle. This could lead to upward migration of volatiles and lighter, more siliceous components which, during the passage through the upper mantle and crust, could acquire by inclusion and partial melting the more basic constituents of andesitic rocks or could lead to the upward migration of a differentiate of andesitic composition.

On the assumption that upward migration of material from deep in the mantle is the primary process involved in the alteration of an area such as this, a possible sequence of developments is suggested. After further differentiation, some process of convective circulation or overturn might cause a large body or many smaller bodies of primary basaltic magma to be collected and to begin migrating upward through the mantle. In addition to an initial density differential, the heat generated during the crystallization of such minerals as olivines and pyroxenes from the primary magma and possible volume increases associated with changes of state may be major driving forces in the migratory process. The rising magma might thus permeate a large area of the upper mantle and crust. Except in the upper regions of the mantle and in the crust, passage of the magma would probably cause little or no permanent change in the penetrated materials, since the rising magma would presumably be differentiating during the ascent, leaving behind the more basic constituents. In the upper regions, however, where cooling would eventually halt the migration, the end product could be a contaminated mantle and crust, which would perhaps have the accumulation of varying amounts of the lighter, more siliceous materials on the surface. In addition, broad areas of sediments might be metamorphosed or cemented by the volatiles that reach the surface. Where vast amounts of the migrating magma rose, such as along the Antilles arc and the Aves swell, the portion with crustal density would have no tendency from buoyant forces to rise above the existing crust, and the result would be a thickening of that layer, as is observed. Isostatic forces acting on such a thickened crustal section would produce the observed elevation of that material. The lighter, more siliceous constituents would constitute most of the upper part of the section, including the islands.

The earliest-rising materials, accompanied by larger amounts of water vapor and other volatiles than those succeeding, may provide the mechanism for the implacement of ultramafic rocks, particularly serpentines, as suggested by Hess (1955, p. 402). This is believed to occur at an early stage in the intrusive process.

Depression in the form of deep-sea trenches probably results from compressional forces. For orogenic belts along continental margins, the compression could be supplied by the tendency of continents to spread into the oceans (Stille, 1955). For an island-arc orogenic belt, the compressional forces may originate from extensive intrusion, as postulated, which contaminates and expands a large area of the upper mantle and crust. As a result of this expansion, laterally and vertically, the tendency at the borders will be for the altered interior basin to move out over the exterior oceanic basin. Such a stress distribution could account for the down-buckled deep-sea trench and for the major fracture zone of earthquakes dipping down under the island arc, as suggested by Gunn (1947). Along the raised ridge of the island arc local relaxation of compression in the upper portion of the mantle and crust will provide a zone of weakness along which volcanic activity would be concentrated.

Visualization of island-arc orogenesis resulting from differentiation and upward migration

of volatiles and the more siliceous materials from the deep mantle must be concerned with the problem of concentration or focusing of energy. Why does this process occur in the neighborhood of continents only and never in the middle of an ocean? Any hypothesis of accretion and growth of the continents in place must imply localization or focusing, throughout geologic time, of the processes involved in collecting the more siliceous, lower-density materials. Thus a recent migration might follow the pattern of the past and occur under a continent or near a continental boundary. Further intrusion under a continent would produce continental growth by thickening, uplift, and subsequent spreading. Extensive intrusion under the ocean might produce a result similar to that observed in the Caribbean, *i.e.*, alteration and expansion of a broad region of the crust and upper mantle; the expansion would supply the tectonic forces which produce the island-arc—deep-sea-trench structure.

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LAMONT GEOLOGICAL OBSERVATORY (COLUMBIA UNIVERSITY), PALISADES, N. Y.; THE RICE INSTITUTE, HOUSTON, TEXAS, AND WOODS HOLE OCEANOGRAPHIC INSTITUTION, WOODS HOLE, MASS.; WOODS HOLE OCEANOGRAPHIC INSTITUTION, WOODS HOLE, MASS.; WOODS HOLE OCEANOGRAPHIC INSTITUTION, WOODS HOLE, MASS.

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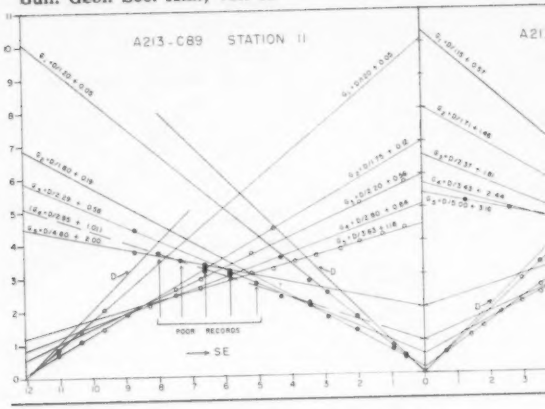
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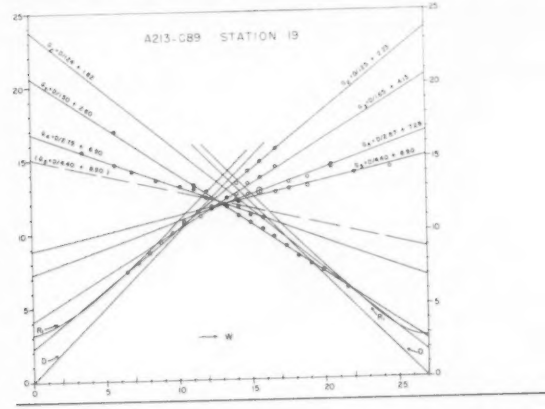
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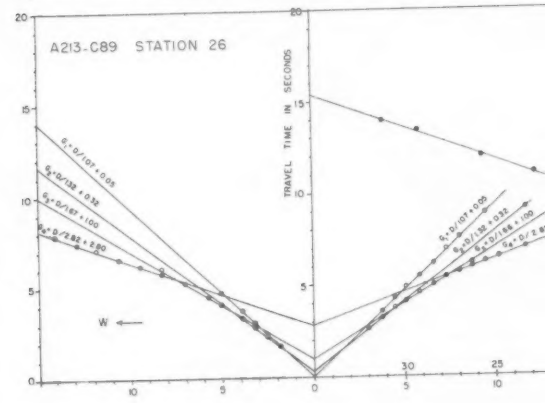
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TRAVEL TIME IN SECONDS
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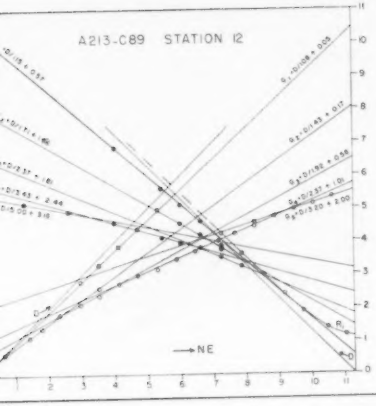
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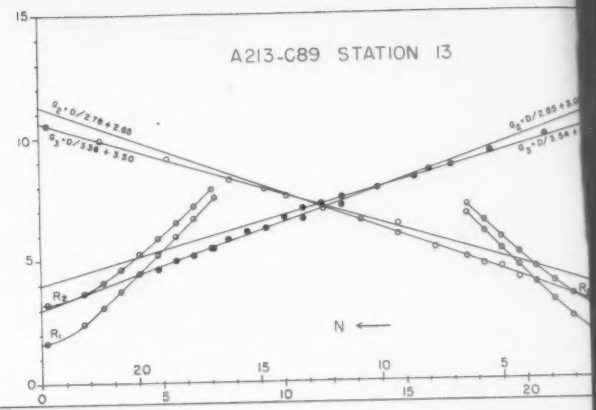
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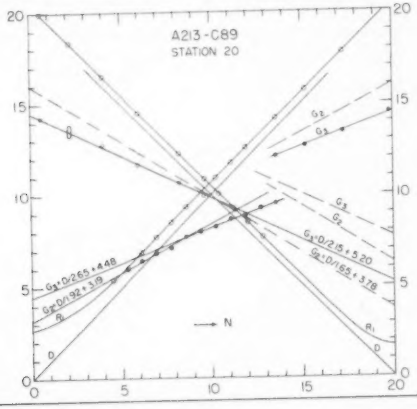
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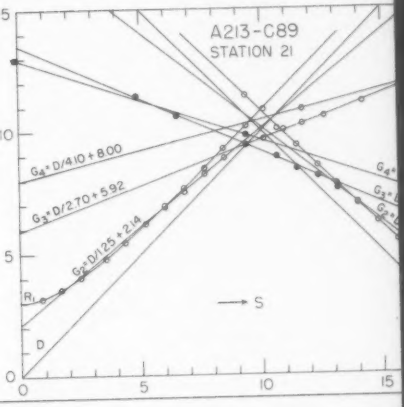
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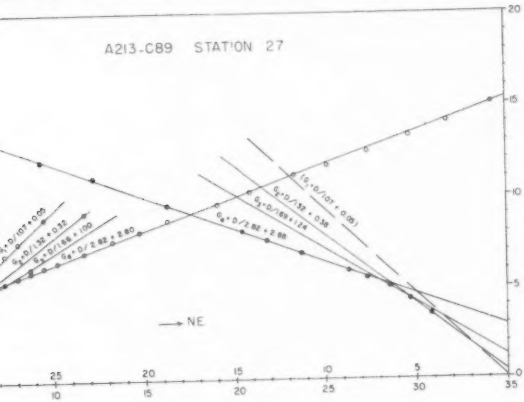
S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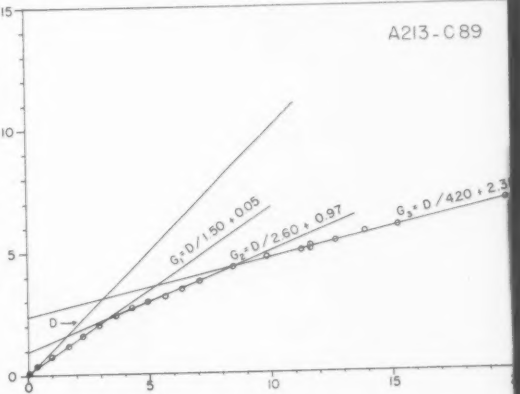
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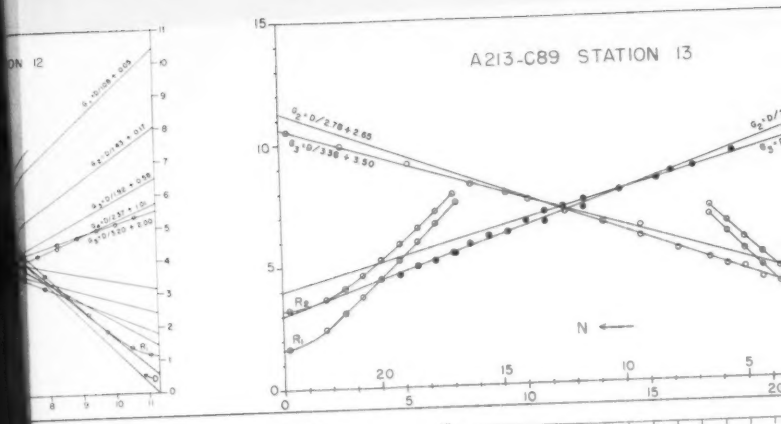
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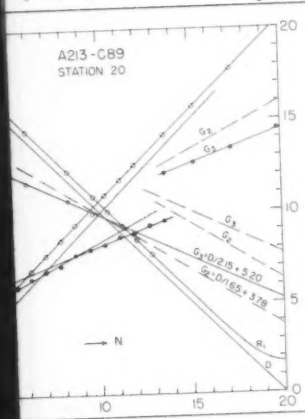
G_2=D/2.78+2.65
G_3=D/3.36+3.50
R_2
R_1
N ←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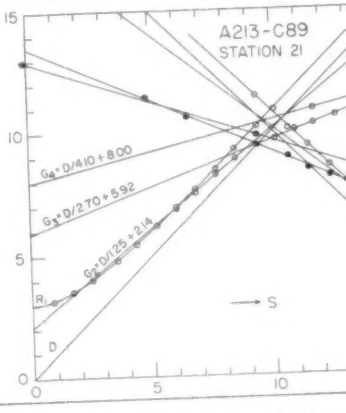
STATION 20
G_2
G_3
G_4
G_3=D/2.15+5.20
G_2=D/1.65+3.78
R_1
D
→ N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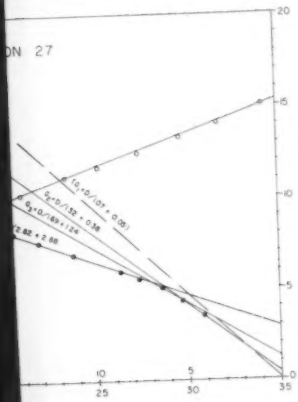
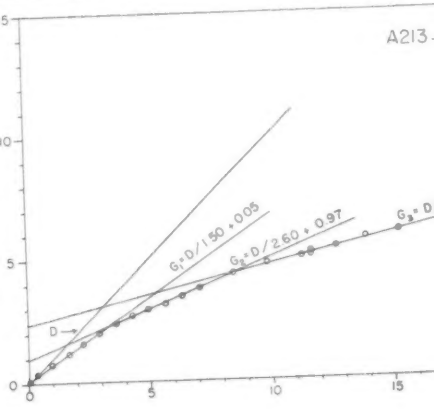
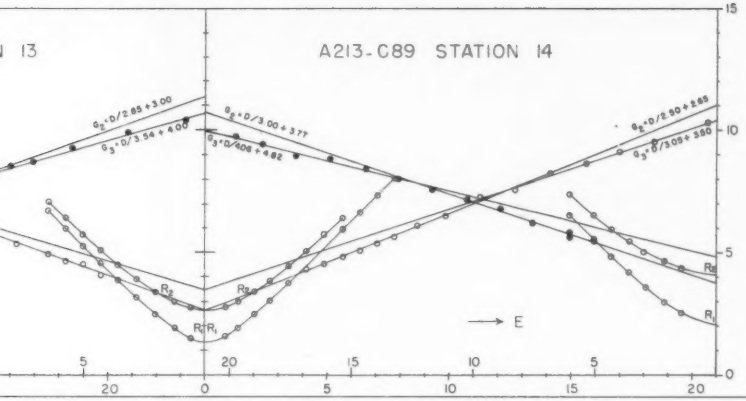
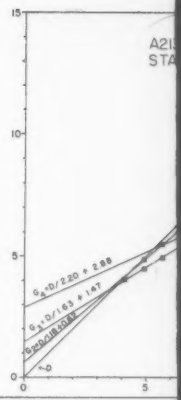
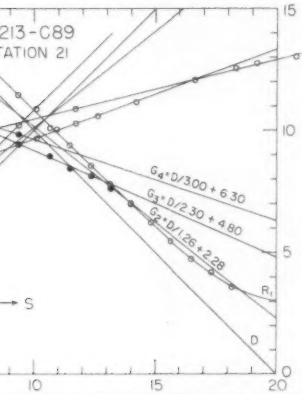
STATION 21
G_4=D/4.10+8.00
G_3=D/2.70+5.92
G_2=D/1.25+2.14
R_1
D
→ S

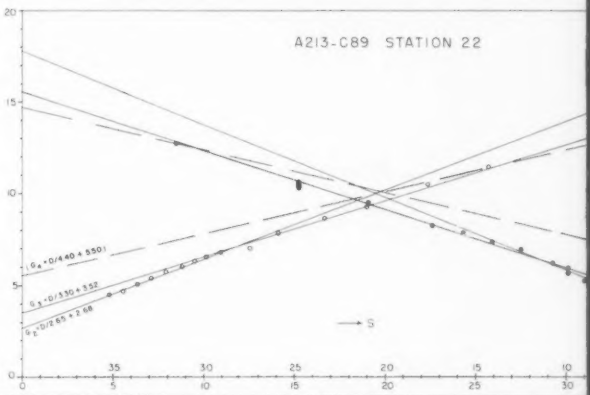

ON 27
G_3=D/1.32+0.98
G_2=D/1.69+1.26

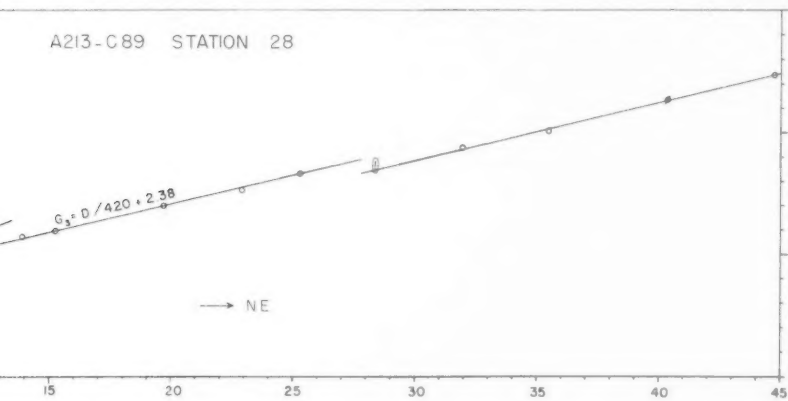

A213-
G_1=D/1.50+0.05
G_2=D/2.60+0.97
D

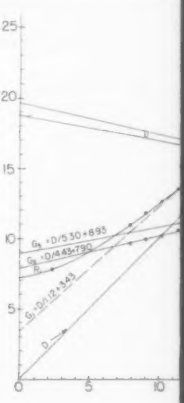






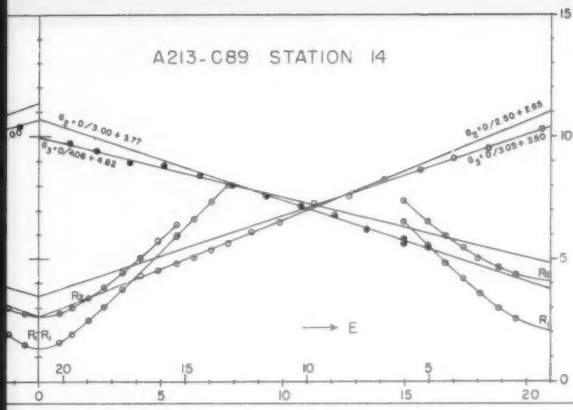


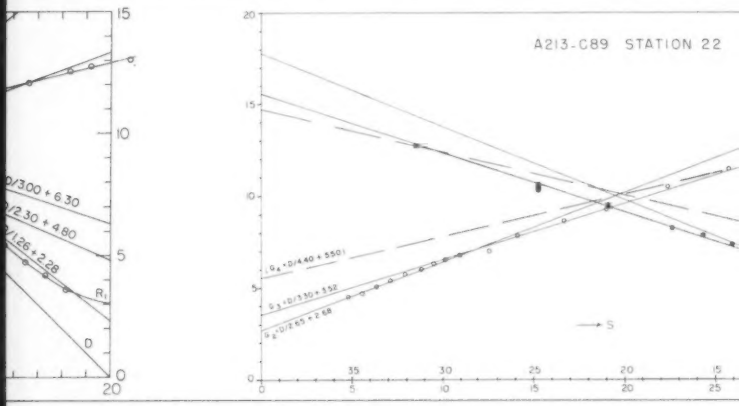


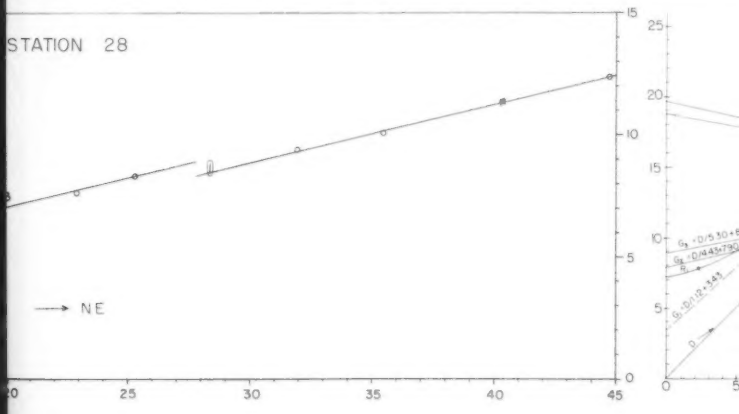


TRAVEL-TIME G

Stations A213 to C89 (nos. 11-28) and stations G-5 to G-8. Distance on the abscissa is plotted in increments of direct water-wave (d) travel time, *i. e.*, 1 second is approximately equal to 5030 feet. The lines G_1, G_2, and G_3 are determined by refracted arriva the o the t

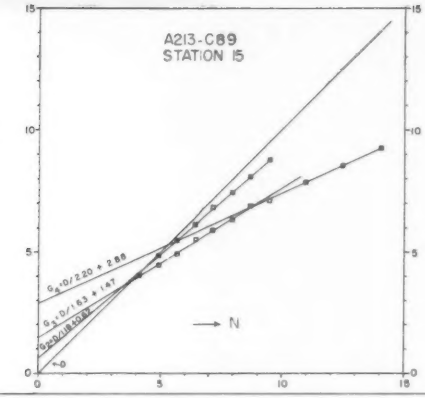


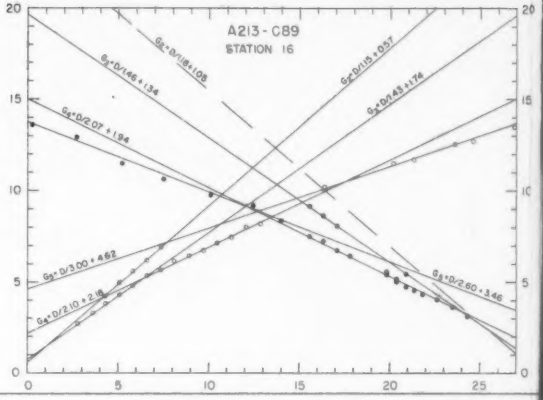


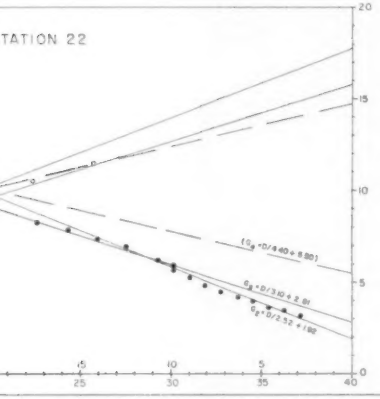


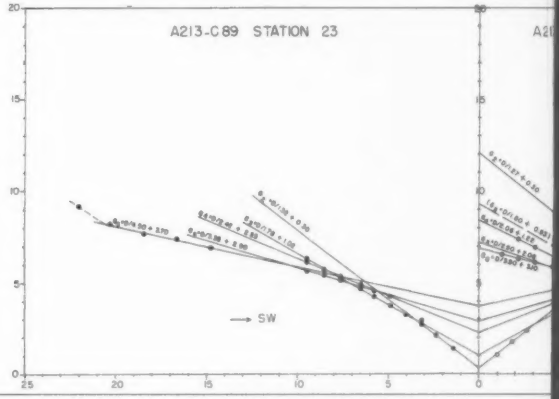
TRAVEL-T

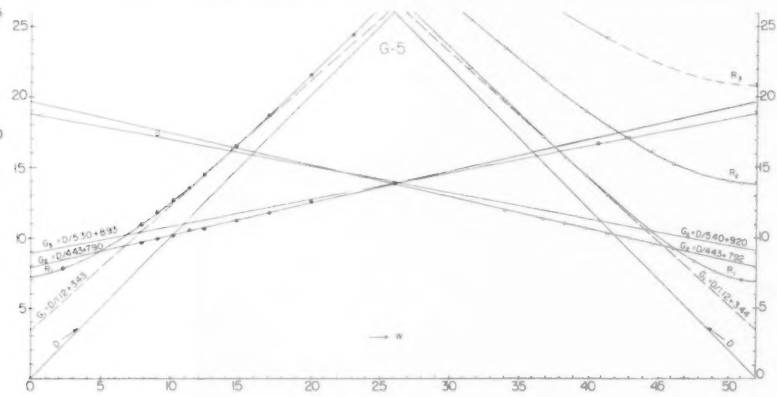
Stations A213 to C89 (nos. 11-28) and stations G-5 to G-8. Distance on the abscissa is plotted in increments of direct water-wave (d) travel time, *i. e.*, 1 second is approximately equal to 5030 feet. The lines G_1, G_2, and G_3 are determined by refracted

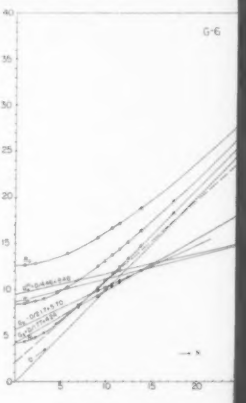






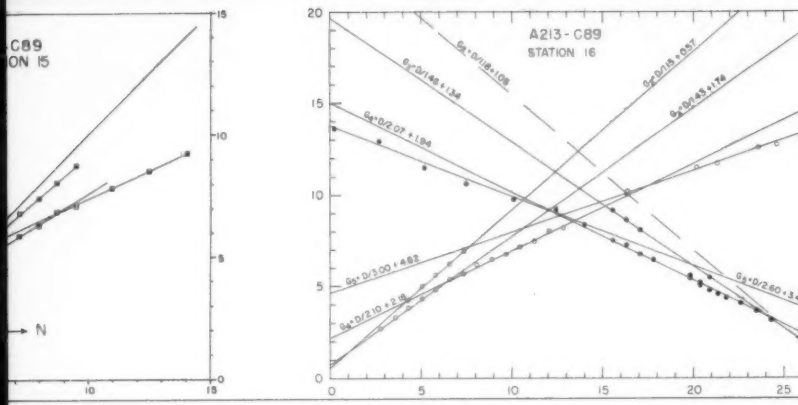


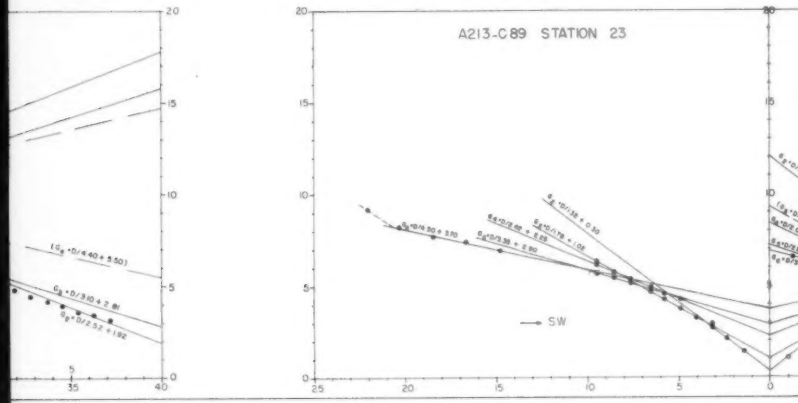


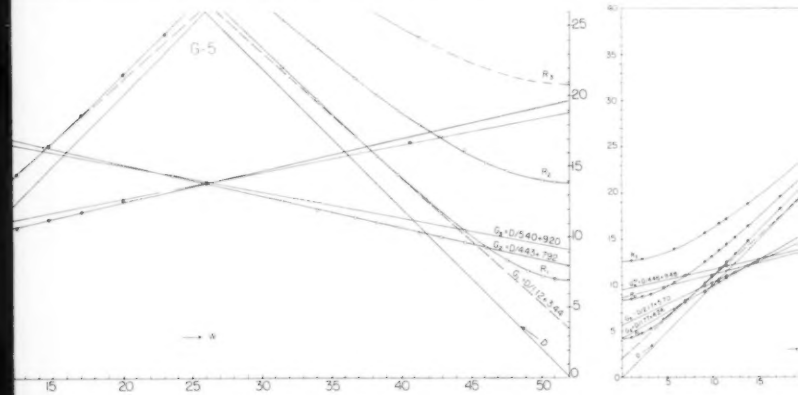


AVEL-TIME GRAPHS

-5 to G-8. ect water- equal to refracted arrivals traveling in the first, second, and third layers beneath the ocean floor. At stations where no G_1 line appears, velocity in the top layer was assumed.







RAPHS

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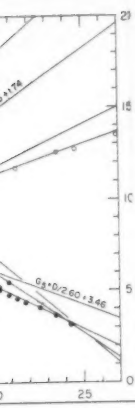

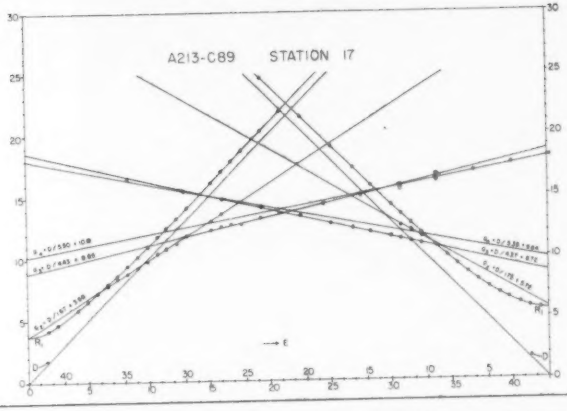
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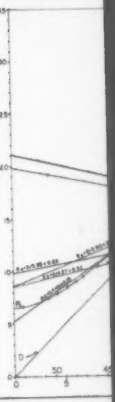

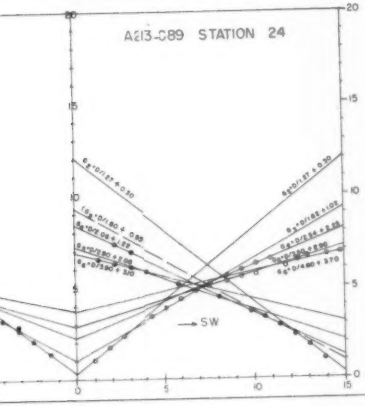
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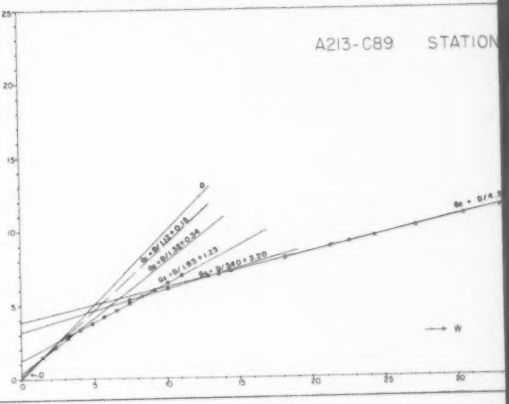
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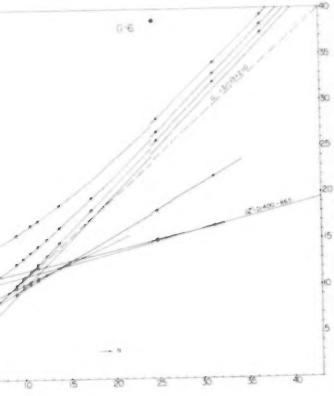
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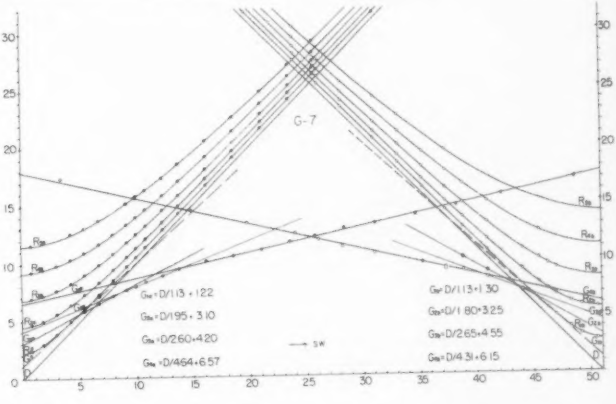
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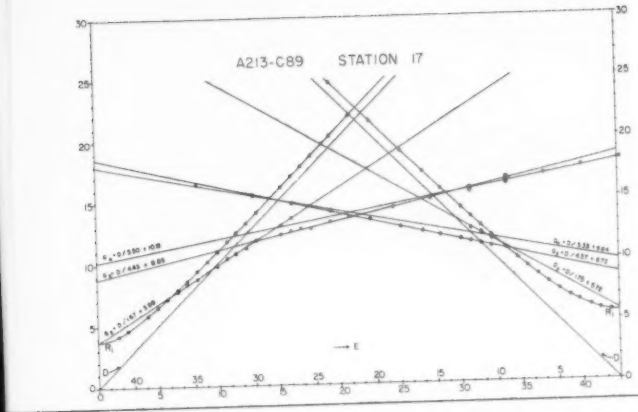
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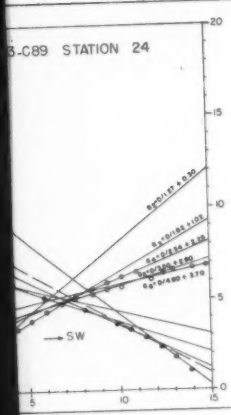
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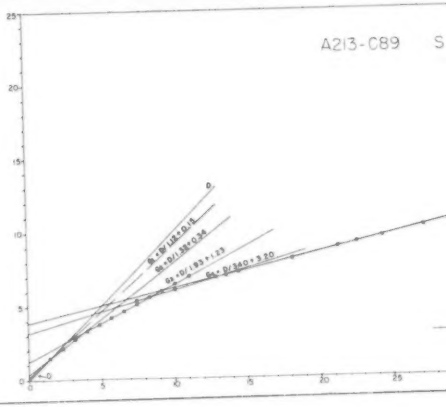
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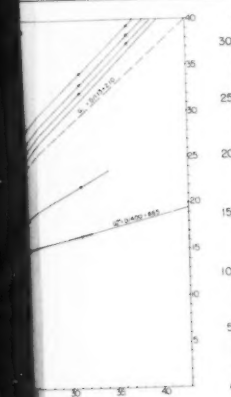

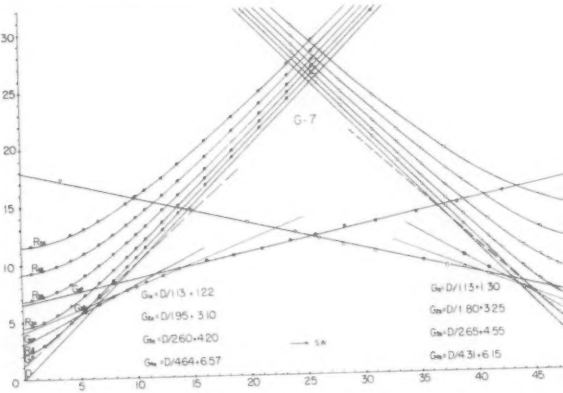
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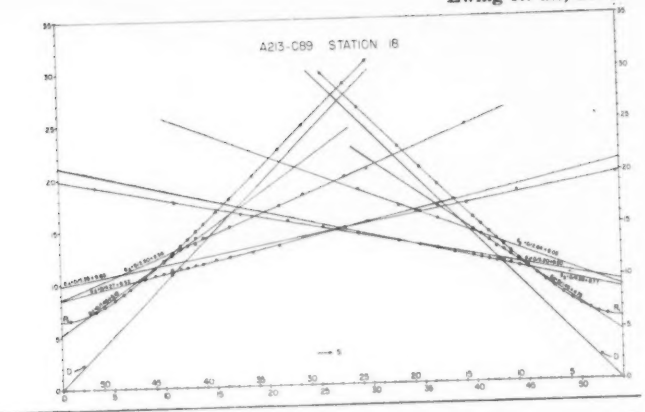
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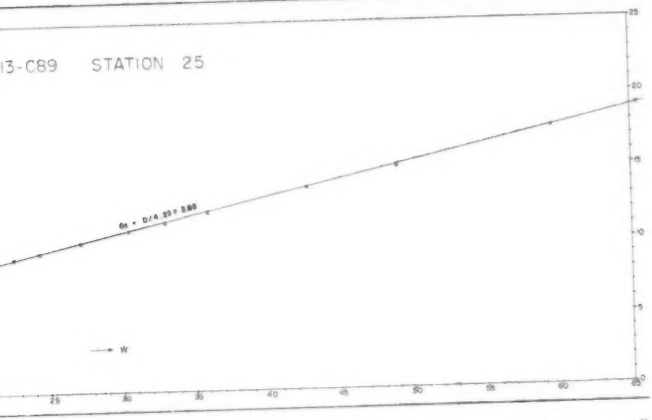
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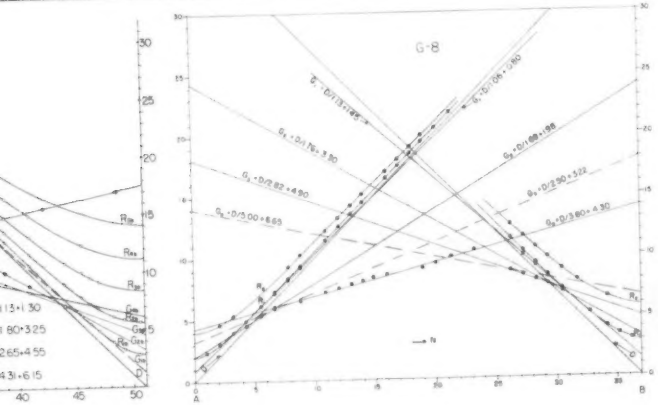
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BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 68, PP. 913–930 JULY 1957

PENEPLAINS AND PEDIPLAINS[1]

BY HENRI BAULIG

Translated from the French by C. A. COTTON

ABSTRACT

(Prepared by the translator)

Professor Baulig agrees with L. C. King in that he adopts as probably correct the theory that pediplanation has produced all the ancient planed land surfaces of continental extent which truncate even the most resistant rocks. He rejects, however, King's claim of universality for pediplanation and affirms belief in the process of downwearing of the land surface under humid climates. He claims, that planed surfaces on uplands in regions now temperate originated as true peneplains during a widespread late Tertiary extension of the hot-humid, or "tropical", climatic zone. He is critical of a number of the assertions included by King in the "canons" of landscape evolution, advances substantial arguments against some of King's claims for strictly parallel backwearing, and counters the bold statements that all surfaces hitherto called peneplains are pediplains and that the peneplain is "an imaginary landform". He also rehabilitates the concept of normal erosion and justly claims that no acceptable word has been proposed as a substitute for "normal" in this connection. He champions the principle of the profile of equilibrium.

CONTENTS

TEXT

	Page
Introduction	913
Parallel retreat or reduction of declivity by downwearing?	916
Pedimentation	921
Pediplains or peneplains?	923
References cited	928

INTRODUCTION

Modern geomorphology has developed for the most part in countries the climates of which are average in the sense that they are neither very hot nor very cold, neither very humid nor very dry; they may be described as humid-temperate.[2] Davis, who made a great contribution to the progress of this science, spoke of "normal" climate, cycle, forms, and processes, as distinguished from arid and glacial "accidents". However, attention has been directed increasingly toward the relation of relief forms to climate; and as a result areas, each of which is dominated by either a single process acting almost to the exclusion of others have been recognized, such as erosion by wind or frost, or, more usually, combinations, though in variable proportions, of the same primary processes—by erosion systems. Every such more or less clearly defined region has thus now, and has had in every past geological age, its "normal" regime, controlled by the climatic conditions now or then prevailing.

"Normal" in the sense in which it was used by Davis is certainly wrongly applied, but this incorrect usage may be tolerated. There are other examples of similar misuse in scientific vocabularies, for example the meteorologist's calculation of the "normal" temperature, pressure, and humidity of a place, and the hydrologist's use of "modulus" with the meaning of mean values only, though these are not necessarily the most frequent; just as in geo-

[1] This paper has been translated from the text in Bulletin de la Société Belge d' Études Géographiques Tome XXV (1956), no. 1, p. 25–58 (four line figures and the Appendix, p. 54–58, omitted).

[2] Nevertheless, as Davis remarked, most of the fundamental concepts developed by Powell and Gilbert in the semi-arid West of the United States are quite applicable to humid morphologies.

logical usage "normal" faults are no more normal than are reverse faults. The use of "normal" can be justified, however, only because no substitute that is both exact and convenient has so far been found for it; "humid-temperate" is an awkward expression, and "fluvial" and "pluviofluvial" are just as correctly used of regions that are neither temperate nor humid. All things being considered it is not regrettable that certain forms should have been described as normal, for it is as compared with these "norms" that the "anomalous" nature of glaciated valleys in the Alps and the "anomalies" of arid and in particular of subarid regions have eventually been recognized.

Hitherto discussion has concerned only terminology, which is a minor matter. However, a vigorous assault has been made on the idea rather than on the word by Lester C. King who affirms that substitution of "normal" for "humid-temperate" is not only an impropriety but a usurpation. The "normal" morphology, if there is one, King claims, is (and has been in the geological past) that of semi-arid regions, which he (1951; 1954) has described in the notable work *South African scenery*.

South Africa, as has long been known, is a particularly stable segment of the earth's crust. Consolidated early in the Precambrian, it has since, except for the Cape folding, undergone no deformation other than warping into large basins and broad swells, marginal flexure, and the very recent fracturing in the rift-valley region. Since remote times the subcontinent has, except for a narrow littoral zone, escaped marine transgressions; the ancient basement, mainly crystalline and metamorphic, is overspread only by terrestrial sediments, which are thick, however; they are those of the Dwyka, Karroo, and Kalahari systems.

The most characteristic features of the morphology are vast planation surfaces dating from several periods, Precambrian and later, some of them unburied, others temporarily covered but again exhumed. The surfaces intersect, but commonly at very small angles; new covering beds are deposited without much erosion of the basement rocks. Thus the ensemble is complex; but the more recently developed surfaces step down regularly from one level to another, each corresponding to a major erosion cycle—Cretaceous, Miocene, and more recent cycles (King, 1951, map at end). In the opinion of King, however, these surfaces are not peneplains developed (according to the Davisian theory) to some extent by enlargement of the floors of valleys but to a greater extent by gradual reduction of the steepness of hillslopes; rather are they coalescing pediments—pediplains—such as have been developed as hillslopes have retreated and owe their formation to "sheet wash". Each major cycle has been initiated by an accentuation of the continental marginal flexure, which has caused a wave, or several waves, of erosion to make their way headward along the courses of the main rivers and then, by degrees, up those of their tributaries. Pedimentation, closely following valley incision, has enlarged the valley floors, with the result that the interfluves have been eroded away and have been reduced eventually to isolated hills ("inselbergs"), to kopjes, to mere nubbins, or tors. Well before a pediplain can be fully developed, however, a new cycle, originating likewise at the continental margin, has begun to attack the already planed areas, so that two or more cycles successively initiated have been developing simultaneously, each encroaching on the territory of the preceding one and yielding ground to that which follows it.

These conclusions have been reaffirmed in a recently published article (King, 1953), in which they are used to the extreme, extended in their application to almost the whole earth and its past history, and finally condensed into 50 "canons". Perhaps the most significant of these are No. 3, 19, 24, and 37.

Canon 3: There is a general homology between all epigene landscapes. The difference between landforms of humid-temperate, semi-arid, and arid environments are differences only of degree. Thus, for instance, monadnocks and inselbergs are homologous.

Canon 19: The pediment is the fundamental landform to which epigene[3] landscapes tend to be reduced the world over.

Canon 24: The standard, or "normal", type of landscape, both now and in the geological past, is the semi-arid type with broad pediments and parallel scarp-retreat.

[3] "Epigene" means subaerial exclusive of glacial and karst processes.

Canon 37: A peneplain in the Davisian sense, resulting from slope reduction and downwearing, does not exist in nature. It should be re-defined as an "imaginary landform".

Without pausing to remark on the trenchant wording of these assertions and of many others that accompany them, let us examine whether propositions probably valid for the greater part of South Africa are equally so for the rest of the world and for all geological periods.

The possibility of planation by the progressive wearing down of hillslopes has been in doubt almost ever since the idea was conceived, for the efficiency of the erosive processes, a direct function of the slope, must decrease as the slope becomes less steep. That is why Davis himself thought that planation could scarcely proceed beyond the peneplain stage on resistant rocks on which only an undulating surface with a local relief (except for manadnocks) of several hundred feet was developed. There is no doubt, however, that surfaces have been more thoroughly planed than that by subaerial processes (Baulig, 1952a).

Powell (1876, p. 196), who was aware of this difficulty, considered that solution weathering, which must be active even on slopes already reduced to small declivity, could complete the task. King (1953, p. 722) doubts this, for, as he says, the planed surfaces with which he is familiar have soil covers that are not at all impoverished by leaching but are calcareous, siliceous, and ferruginous duricrusts which indicate return to the surface of water of infiltration bringing with it dissolved substances. In the humid tropics, nevertheless, solution is known to be extremely active and, as Freise (1938) has shown, attacks all minerals. Moreover, measurements of the turbidity and of the salinity of river waters have established that chemical denudation accounts for at least half of total denudation throughout the United States and for a still greater proportion in the humid parts (Baulig, 1910).[4] King's view that the existence of a residual, or eluvial, soil is not in itself a proof of peneplanation is correct, because after a change of climate such a soil can be formed on a surface that has been developed under quite different conditions; and the converse of this is no less true. Further, the soils of humid climates, usually poorly consolidated, are much less likely to be preserved than are concretionary materials formed under semi-arid conditions, and a paleoclimatology based entirely on pedology would be in danger of missing most of the humid phases (Baulig, 1952b, p. 231).

King emphasizes another difficulty encountered by the peneplanation theory. This cycle is characterized by rather early appearance of maturity of relief, *i.e.* of almost complete dissection, but, if increasing slowness of cyclic development is assumed, the attainment of old age, or peneplanation, is very long delayed. Erosion in a second cycle must, therefore, have eliminated forms of the first cycle long before the land surface had been reduced in that cycle to a peneplain. Planations that afford evidence of successive cycles can exist at the same time and but a short distance apart, but they occur on terrains that are definitely unequally resistant to erosion. King, however, in agreement with other authors, records that in South Africa downstepped surfaces cut on very similar rocks are common, and also that inselbergs are not generally different lithologically from pediments they surmount.

The objection to the peneplanation theory that is thus raised is a strong one if, in agreement with Davis, it is believed that new cycles are introduced by warping or differential upheaval which disturbs equilibrium established in a preceding cycle and ends that cycle. If it is believed that successive cycles are started by lowerings of base level or by uniform upheavals the objection is less strong. In the latter case each cycle, propagating itself gradually towards higher ground, may reach an advanced stage farther down valleys without as yet destroying all traces of earlier planations. This concept is not a new one; it is applicable rather to local peneplains of moderate extent than to great planed surfaces extending over a great part of a continent. Such surfaces are, however, rarely monogenetic; it is demonstrable that some and

[4] Though cultivation of the land and overgrazing during the last 2 centuries have certainly promoted soil erosion and thus increased the solid load of rivers, chemical denudation has not been affected to the same extent. It follows that the share of chemical processes in denudation was proportionately much greater in the past.

it may be assumed that most other surfaces are composed of facets of different surfaces of planation that have been developed successively and intersect one another; some of these have remained buried and, though recently exposed, have been preserved for a very long time. Such is the case in South Africa as well as elsewhere. Furthermore, King's major cycles are really composite: he points out that local warping may lead to revival of erosion and thus initiate a subcycle which is propagated for perhaps a considerable distance. Should a wave of headward erosion expose an outcrop of resistant rock such as one of the dolerite sheets interbedded in the Karroo, the wave splits in two; erosion, which was arrested or at least retarded by the obstacle, can continue without check on the overlying weak beds (King, 1951, p. 174, 241). Notwithstanding these reservations, King's conclusions as applied to South Africa and similar regions may be considered valid without admitting the unlimited extension he claims for them.

Parallel Retreat or Reduction of Declivity by Downwearing?

King's thesis is based essentially on the postulate that after hillslopes assume a certain declivity characteristic of rock and climate, they maintain this steepness unchanged throughout their erosional history, at least as long as they retain some topographic importance (1951, p. 46; 1953, p. 723, Fig. 1): there is no such thing as "downwearing", but only "backwearing". Davis, on the other hand, and those who follow him have thought that whereas a rock slope can under certain conditions maintain a sensibly constant inclination as long as bare rock is exposed on it, a slope that is covered tends to be lowered, provided the local base level remains approximately unchanged. (By "covered" is meant that it has on it a "soil" in the geomorphological sense: it is overspread, that is to say, by a mantle of debris that has resulted from disintegration of the bedrock of the slope itself and is held in place by a sufficiently dense vegetation.)

If a hillslope that is attacked incessantly by destructive processes maintains an almost constant declivity, a balance is clearly being maintained between the production and the removal of debris. At any particular time, however, these both depend on the nature of the rock and on the climate: if the rock is practically homogeneous and if the climate remains approximately constant one can see no reason why the slope should become steeper or less steep. Abundance of debris and also the sizes and shapes of constituent fragments, on which the mobility of the debris depends, are controlled mainly by the jointing in the parent rock and the susceptibility of the rock to both mechanical and chemical weathering, which depend on climate. Wastage by removal of debris is in the last analysis effected by gravity acting either directly (talus streaming, slumping, landsliding) or through the agency of rain wash, over which there is some climatic control. The mere fact that a hillslope remains a surface of bare rock is a sufficient proof that all disintegrated material is removed from it as it is produced, and probably the processes effecting removal could dispose of more such material if it were available.

Some of the steepest slopes are on the most resistant rocks (*e.g.*, quartz veins) regardless of climate and destructive processes; more commonly they are on rocks that are very resistant as long as they are fresh but lose their coherence when weathered. Isolated hills in tropical regions (inselbergs) attain their steepest average slopes (50° and more) in crystalline rocks (such as granite) that are subject to granular disintegration and under a climate with dry seasons, which are long and very dry, so that dense vegetation is interdicted; these alternate with seasons of rather abundant rains, which favor weathering of rocks and quick evacuation of debris.

Slopes are less steep under the most arid climates, where the dryness sets a limit to comminution of debris. Debris lingers on the slopes, and the hillslopes are "half-covered". If the rock is such that it furnishes angular fragments of various sizes that resist comminution, the removal of waste takes place slowly and the inclination of the hillslope becomes moderate; debris also accumulates temporarily at its foot as a talus slope, though this is gradually degraded. If, on the other hand, the rock, granite for example, breaks at first into large blocks and these crumble later into sand, rain wash easily sluices away the sand and

leaves the blocks; these maintain a much steeper inclination of the hillslope than in the preceding case. The blocks break down eventually, however, into sand, either on the slope or at its foot, after which rain sweeps everything away cleanly.

In the foregoing it has been tacitly understood that the slope is not undercut by vertical stream corrasion or sapped at the base by lateral stream corrasion (meander cutting, for example), by wave attack, by wind, by glacial erosion, or finally (as described below) by pedimentation. In these eventualities the wall-like slope will be at least partly oversteepened, and a scree may accumulate at its base. If the scree becomes stable (as at the foot of an abandoned sea-cliff, in the amphitheater of a cut-off meander, or below the wall of a glaciated trough), the hillslope, which will be progressively buried at the base, will continue to recede with constant declivity above the basal accumulation, whereas the rock face under this will tend to assume a convex profile decreasing in curvature upward.[5] Except for these usually temporary conditions, in rock that is nearly homogeneous and under a climate that varies but little a lasting equilibrium is established; it may be described as static because the ratio of forces to resistances is practically constant, that is, it does not change in one direction as time goes on.

The same is not true of the slope "with cover" (Baulig, 1940; 1950, p. 125–147). If, because of the lithology, climate, vegetation, and stage of cyclic development, the products of rock disintegration are not progressively removed they tend to remain as an overspreading sheet at first discontinuous, thin, and of coarse texture, but afterwards progressively more continuous, thicker, and finer in texture. From now on the rock is no longer directly exposed to atmospheric weathering but is weathered only beneath the enveloping waste mantle, the "regolith" (Merrill, 1904, p. 299). Though this affords some protection, because it is mobile it is always being renewed at the expense of the underlying rock. There is a tendency, therefore, for an equilibrium to be established in which, except for the unvarying factors rock, climate, and vegetation, a number of interdependent variables take part. These are thickness, texture, permeability, and mobility of the waste mantle on the one hand, and on the other the strength of the agents responsible for removal of debris (ablation)[6] which for the most part depends on the angle of slope. A profile of equilibrium, like that of a water stream, can re-establish itself spontaneously if it is upset and also continues to readjust itself throughout the cycle. In terrains that are approximately homogeneous (at least as regards the ensemble of rocks) a hillslope that is fully mature exhibits a compound curve, convex on the upper and concave on the lower part of the slope. The convex segment is shaped by creep[7] and by rain wash that is as yet so feeble that it is likely to be stopped by the smallest obstacle and is thus described as "unconcentrated"; it remains discontinuous and diffused. Steepening of the slope downhill is a result of the increasing effect of gravity on an increasing mass of material with small mobility. Concavity of slope, on the other hand, results mainly from the activity of a more abundant runoff now gathered into rills which pick up and carry off finely divided particles from materials that are becoming progressively less permeable and more easily removed. Ephemeral as it may be, concentrated rain wash acts continuously, at least for short distances. There is, therefore, a tendency to develop the concave longitudinal profile charateristic of running water, and the incipient rill channels or gutters impose their own profile, howbeit with reduced curvature, on the lower slope as a whole, as they gather in from it the runoff and carry away the debris. The relative extent of convexity and concavity varies with humidity of climate and permeability of the rock, or, more correctly, of the debris furnished by the rock (*e. g.*, granite sand), and this proves that the processes dominant on upper slopes decrease in their relative value downhill,

[5] This theory was developed first by Osmond Fisher (1866) and later, much more fully, by Lehmann (1933).

[6] Translator's note: For a definition of "ablation" *see* Baulig (1956, paragraph 5).

[7] Creep is not, properly speaking, a mass movement. It is simply the sum of an infinite number of elementary displacements of particles with respect to one another, which, as gravity is constantly intervening in the same sense, causes movement of the whole mass. In landslides, mudflows, and solifluxion it is on the other hand movement of the mass as a whole that determines relative displacements of the material composing it.

and vice versa. Also, since the transition is gradual, the passage from the convex to the concave slope is made at a point of inflection or through a neutral rectilinear segment.[8]

All parts of such a profile are interdependent, since the declivities at all points, which make up the curve, do not individually depend merely on what happens at these particular points but also on the contributions of water (from runoff or seepage) and waste material that are brought from higher up the slope, and also on their downhill evacuation. Such is the true concept of grade, of which, however, King (1953, p. 725) discounts the fundamental value.

The profile of equilibrium is mobile at every moment and throughout the whole cycle. Obviously the overall gradient of the hillslope, which is constantly losing material and which is under the control of a local base level assumed fixed in position, is reduced in steepness. At the same time, however, its profile tends to be modified, as follows. On a gentler slope the waste, moving downhill more slowly, is further reduced in fineness of texture, so that concentrated rain wash can more easily take over transport. Thus the concave segment extends at the expense of the convex. The convex segment, on the other hand, shaped by weathering, creep, and unconcentrated rain wash, processes the intensity of which scarcely decreases with reduction of altitude, can preserve the initial curvature in the part of the curve that survives, though the length of this segment is progressively reduced. In an extreme case, in the superhumid tropics, the waste mantle, though thoroughly decomposed by weathering and saturated with water, is held in place by tree roots, but only until patch by patch it slides away and takes with it its cover of vegetation. The concave segment can then extend to the top of the slope, so that the valley assumes a catenary transverse profile between sharp-crested interfluves.

To this well-founded concept King opposes one of his own. Instead of discussing homogeneous structure first and reserving more complex cases for attention later, he describes a tabular horizontal structure comprising, like the Karroo, intercalations of very hard rock, dolerite, in an alternating succession of sandstones and shales that are not on the whole very resistant (King, 1951, p. 45; 1953, p. 728). The landscape profile[9] is made up of four distinct elements: (1) the escarpment ("free face"), a cliff corresponding to the vertical section of the hard stratum; under the attack of weathering and gravity it is diversified only by infantile ravines; (2) above, in weak rock, the "waxing" slope, a description justifiably criticized by Cotton (1948, p. 227) as implying a particular theory of development of slopes which is not in question (convex slope may be suggested as more appropriate); it is shaped essentially by creep; (3) below the cliff the scree (the "debris" or "detrital" slope), veneered with debris fallen from the cliff the coarseness of which determines the declivity; (4) lower down, the "waning" slope (objection the same as for "waxing"), to be called otherwise the concave slope; which passes into the pediment. These four elements King claims to find on every hillslope undergoing active erosion, more particularly the "free face" and the "debris slope", which retreat with unchanging declivity. Further, he extends this claim of parallel retreat to the whole slope, though it is true only of the escarpment, if one is present, and of the upper part of the scree.[10]

King (1953) does admit exceptions to his thesis, exceptions which largely destroy its value as a generalization. Thus (Canon 4): "Four elements may occur on a hillside slope . . . Each or any element may be suppressed . . . Canon 6: "If [the free face and the debris slope] are actively eroded the hillside will retreat parallel to itself"; unless (Canon 9) "the waxing slope becomes strongly developed and may extend down to meet the waning slope"; but then the profile is degenerate or "atrophied" (p. 736). (But have the escarpment and the

[8] Here, at all stages of development, is found the steepest part of a convexo-concave slope. Commonly it is the only part that is envisaged by those who argue for parallel retreat of slopes in all cases.

[9] *See* block diagram (King, 1951, p. 45); which is reproduced by Cotton (1948, p. 226).

[10] King (1953) seems to consider the words "scarp" and "hillslope" (or "hillside") as synonymous. Useful discussion is almost impossible without good definition of terms. He writes (p. 723) of "parallel scarp-retreat" and (p. 733) of "retreat of the steeper element of the hillside", but states (p. 728) that "the hillslope retreats with constant declivity", and writes (p. 732) of "parallel retreat of hillsides and scarps", further (p. 736) even admitting the possibility of "*scarp*-retreat under a sod mantle", which is contradictory.

scree slope ever existed, even on resistant rocks, if the adjacent river cut down slowly enough to allow the hillslope to be kept continually graded?) Canon 13: The "old deduction of continuous lowering of hillside gradients . . . is incorrect". Such lowering has taken place only on "terrains of rocks so weak that they cannot maintain a free face and detrital slope". (Is it a matter only of resistance and not at all of structural differentiation and stage of development?)

Even in horizontally bedded structures, like those which King considers, where thick resistant and weak beds alternate, the escarpments and scree slopes will not behave in the same way. The escarpments recede with an angle of slope that is almost constant, but recede with varying velocities, for the rate of retreat depends partly on their own resistance and partly on the rate at which the scree slope below is degraded, a rate which is determined in its upper part mainly by the dimensions of the blocks derived from the escarpment above (here assumed to be of practically constant caliber). The debris tumbling down the slope becomes broken, however, into smaller blocks (the farther it travels the smaller are these); and thus screes assume concave profiles. The length of scree travel varies, moreover, with the horizontal distance that separates the upper from a lower escarpment, the top of which is the base level for the scree slope. If the upper retreats faster than the lower the scree becomes longer, and its slope (except for the upper part) becomes less steep. If, on the other hand, an escarpment retreats faster than those overhanging it, it eventually effaces itself, and two scree slopes become one.[11]

All parts of the profile are thus interdependent; the lowering of the scree slopes depends on retreat of the escarpments, and the converse is also true. The whole hillslope may be said, moreover, to be overswept by an intermittent flood of waste which is being perpetually reinforced, which clears cliffs with a bound, but which lingers on the scree slopes (Davis, 1899).

A profile that shows evidence of one special and persistent structural differentiation will long remain broken, perhaps until the last butte disappears. On a homogeneous rock in which density of jointing varies from place to place a hillslope will, on the other hand, be diversified, and show some steep faces and elsewhere benches on which regolith accumulates. In a humid climate such debris is anchored, or at least partly held in place, by vegetation; and it thus tends to accumulate in a progressively more continuous sheet which will eventually, if the local base level is constant or nearly so, assume a smooth profile.

In contrast with this, under an arid or a subarid climate the rock is almost without cover, and therefore rain wash is active; but the effects of rain wash are different on the upper slope, where it is concentrated in gutters and ravines, and on the lower slope, where the runoff tends to be diffused because of weaker gradient. As a result a break of slope develops, and a nick (re-entering angle) appears between the upper part, or hillslope in a restricted sense, and the lower part, or pediment. When the profile as a whole is considered, however, whether in an arid or humid climate, there is obvious homology. In both cases the summit convexity, if any, is shaped by creep and to a minor extent by unconcentrated wash. In a humid climate the part of the profile corresponding to the "free face" is that near the point of inflection, where the steepest declivity is. The lower slope is concave in both cases, and is subject to regrading, because it is developed by wash, which though ephemeral is continuous in its activity (in a sense that has been defined earlier); the only difference is that in the one case it is concentrated in rills and in the other it is diffuse. As this difference is due essentially to difference of climate, there must be transitional forms between the arid pediment and the humid concave slope.

It is this transition that is in King's mind when he maintains that processes analogous to pedimentation operate even in a humid climate and under forest, which is true, and that they are accompanied by parallel retreat of hillslopes, which is not the case. His idea is based on an article in which Davis (1930)

[11] A rather long scree slope, especially in a subarid climate, commonly, like a pediment, bears traces of rain wash in enmeshed rills or even a sheet. Whether or not such a slope is called a "pediment" the fundamental process involved in its formation is much the same as that which fashions the concave portion of a humid hillslope, and King himself admits this (1953, p. 737; *cf.* Baulig, 1950, p. 84).

examines homologies between planed rock surfaces ("rock floors") arid and humid. Davis (1930, p. 26–27, 136–138, Figs. 5, 6, 7) points out that in a humid climate and in homogeneous rock (granite) the equivalents of arid pediments may be found. When a river, almost stabilized, has begun to develop a flood plain, here and there "valley-floor strips" appear, which are lateral glacis distinct from both the flood plain and the sloping valley side. Concavity of profile is recognizable on these, though Davis shows in his drawings an angle at the meeting place of concave and convex strips. (But the figure is no more than a sketchy diagram.) As a matter of fact, a break in the profile can really be detected for as long as the upper slope is shedding coarse waste (Davis, 1930, p. 149); but at a more advanced stage the upper slope "will coalesce with the soil cover of the widening valley floor in a smooth concave curve" (p. 141); and again (p. 142) "the convex profiles of the residual hills gradually become concave as they are continued down to the broad valley floors, which slant very faintly towards the streams." Davis, moreover, as Cotton (1955) has pointed out in a discussion, spoke of "recession . . ., retreat . . ., withdrawal of the valley-side slopes" (Davis, 1930, p. 136–137), but emphasized that the slopes were being constantly "worn down" (p. 141–142), and he contrasted the diminishing declivity of humid slopes with the constant declivity of arid slopes (p. 148). The concept favored by Davis is thus diametrically opposed to that of King.[12]

Davis, though he did not, as he confessed, know of any "rock floor" cut on granite in a humid region, could cite nevertheless some "broad valley floors" or "valley lowlands" that are surfaces developed by planation but since more or less dissected. Such are, in England, the subsequent depressions between the Oolite and Chalk cuestas, whereas another is the hollowed-out core of the Weald, and, in the eastern part of the United States, the planed surfaces on Triassic sandstones and shales and on weaker zones of the folded Appalachians. In all these, however, the leveled surfaces are on weak rocks; at higher levels there are residual features sculptured from more resistant terrains, and at the contact with these there is a more or less abrupt break of slope.[13] King (1953, p. 136–137) maintains that in the Appalachians of Pennsylvania the inclination of hillslopes is not always the same as the dip of the beds (meaning apparently that it is less, from which it might be deduced that the land slopes had been lowered); he maintains also that in places the basal concavity ("waning slope") extends onto hard rock (meaning that as slopes are reduced in declivity structural influences are progressively less felt).

Macar (1949, p. 269; 1954, p. 18) believes he has recognized benches in the Ardennes that are analogous to the "valley-floor strips" of Davis, and he considers them to be "slopes of minimum transportation . . . such that the agencies of creep and rain wash are now only just able to continue transporting debris over them without having any longer energy to erode". (Davis says, on the other hand, that such strips are being continuously, though ever more slowly, degraded.)[14] The same question crops up here as for the Appalachians, for the structure is similar: is there a break in the profile, and if there is can it not be explained quite well by unequal rock resistance? It must be conceded that (Baulig, 1950, p. 141, note C) in a terrain with homogeneous structure under a humid climate "processes varying in a continuous way (from the upper to the lower part of a hillslope) cannot produce a broken profile". This seems to be confirmed by profiles the writer has reproduced (Baulig, 1950, p. 129) and also by those of Fair (*infra*). It has always been recognized by the writer that the question demands supplementary information—topographic surveys and geological details.

[12] In the discussion on the development of hillslopes there is perhaps some confusion of words. If on a fully mature humid slope only the steepest part is envisaged, namely that near the point of inflection, it can be said justifiably that, while the slope as a whole is becoming less steep, this particular part of it "retreats", but it must be added that it is at the same time being "lowered".

[13] Davis (1930, p. 145) notes that the essentially crystalline Piedmont belt of Virginia and of North and South Carolina has encroached on the Blue Ridge. It appears that the most extensive planations on really resistant rocks have been developed in climates not humid-temperate (*vide infra*).

[14] Macar adds: "These planation levels are referable . . . to the late Tertiary, for which period there are indications of tropical climate." (*See* also *infra*)

Pedimentation

King offers a number of interesting observations on the mechanism of pedimentation, though these perhaps have less value as generalizations than he claims. The phenomenon is easy to define in general terms. The pediment is fashioned by a runoff which because it is always fully loaded cannot become concentrated in permanent courses and erode vertically, but can pick up loose debris and can in certain cases even corrade. The modalities of the phenomenon can differ notably, however.[15]

In 1877 Gilbert developed the theory of "lateral planation." A stream of water which after debouching fully loaded and at a stable level from a course in the mountains swings laterally back and forth, attacks each of its banks, and deposits alluvium behind it as it swings across. It will thus produce an alluvial fan if accumulation is dominant but a rock fan with a veneer of alluvium if this is not the case. The planation process will be especially efficient if the stream carries hard-rock debris and is eroding a piedmont of less resistant material. This is what is happening at the periphery of the Henry Mountains (Gilbert, 1877), in the Big Horn basin (Mackin, 1937), on the piedmonts of the Rocky Mountains of Colorado (Fogarty, 1951) and of the Little Rocky Mountains, isolated on the Great Plains of Montana (Knechtel, 1953), and even on the sub-Atlas "glacis" (Dresch, 1950; Joly, 1950; Birot, 1950). The process is not necessarily linked with an arid or subarid climate. After alpine torrents of the Glacial Epoch ceased to aggrade they eroded glacis which are closely analogous to the foregoing in the sub-Alpine Molasse (Troll, 1926; 1954); this was because low temperatures had considerably reduced the cover of vegetation and frost-weathered rocks furnished abundant transportable debris. Resistant rocks suffer but little from lateral planation in the mountain courses of streams, though there are exceptions to this rule (Davis, 1900). There must be a sufficiently concentrated "stream": neither sheet wash nor braided-rill wash seems capable of undercutting hard rocks (Davis, 1938). King (1953, p. 731, 734) believes that in South Africa at least the effects of lateral planation are insignificant. Rock fans, he finds, are rare and of small extent in that region. The dominant scarp instead of being broken by wide gaps where rivers debouch through it is in many places rectilinear or only slightly sinuous in plan, and it is transverse to the general direction of river flow; it is thus in no way controlled by the rivers. The pediments, moreover, carry neither traces of large river beds nor significant river deposits. Such conditions can be explained in South Africa perhaps by the fact that there are practically no highland areas there that would concentrate drainage; nor are there any weak-rock piedmonts. It must be remembered, however, that on any very extensive glacis with a low gradient the conical form is practically imperceptible; also lateral planation does not exclude other processes that co-operate with it and supplement it.

A sheetflood, such as was described and interpreted by McGee in 1897 and more recently by Ives (1936) in the same Sonoran region (northwest Mexico and the adjacent part of the United States), is not simply sheet wash but rather an extremely sudden and violent flood; by the Arabs it is termed a "seil". On debouching from a mountain course it spreads forthwith over a plain in a thick and turbulent sheet. The front, convex in outline and several feet high, a "wall of water", advances very swiftly—at the pace of a galloping horse (McGee) or 20 miles in 65 minutes (at the rate of 30 km an hour) (Ives). This powerful wave, carrying debris of all dimensions, some from the mountains and the rest picked up from the pediment, is assuredly a very efficient agent of ablation and transportation, though it erodes unweathered rock but little. King, who (because he has not experienced the requisite conditions for such almost instantaneous concentration of the runoff) has not himself observed the phenomenon, is willing to admit that it is capable of retouching a surface that is already flat, but he doubts whether it will plane one (1953, p. 731). It is true that the sheetflood has been observed in all its violence only on pediments already well developed.

All authors have reported patterns of interlacing channels on pediments. King (1953, p.

[15] A distinction is recognized (Baulig, 1950, p. 80–84, bibliography; 1952a, p. 242–252) between desert and savanna planation. *See* also critical reviews by Howard (1938, p. 47–61).

731) does not regard "rill work" by itself as an efficient agency for planation, because, in his opinion, it tends to become concentrated and will thenceforward cut down vertically instead of planing. Davis (1938)[16] has clearly shown, however, that on a bahada or alluvial glacis (and apparently also on a rock-floored glacis, or pediment, with thin detrital cover) there is an unstable system of channels through which runoff takes place by pulsations. When it becomes concentrated in streams the water accelerates, erodes, and picks up debris; then, with the slightest reduction of velocity, it deposits again, obstructs its course, and spreads out; after which it becomes concentrated again and repeats the process. The pattern of channels resembles an arrangement of very flat cones (fans). The streaming water spreads out on one fan, gathers into channels again at its periphery, and then spreads again over another fan. The channels function merely as such during moderate rains, which are heavy enough to wet the ground and to make it impermeable though not to flood it, and they so function also when floods are subsiding, just as a wave breaking on a sand beach smooths it; but as the wave withdraws the water flows to some extent in definite channels, which the succeeding wave obliterates again.

In King's opinion (1953, p. 733–734) the agency essential for pedimentation (at least in South Africa) is sheet wash (not "sheetflood"), the water for which is furnished by rain that has fallen partly on higher ground and partly on the pediment itself. He has found from observations on the veld during moderately heavy showers (as much as 2 inches an hour) that, though small streams that flow off the escarpment are turbid, on the pediment the sheet flow, which is no more than a fraction of an inch deep, is clear from the inception of its flow—an indication that though flow is turbulent and erosive on the steeper slope it becomes laminar and non-erosive on the platform at its foot.[17] At this point there is a hydraulic discontinuity (which must cause some deposition, though the deposits will be removed at some time by more vigorous wash). Farther down the pediment slope the flowing sheet of water, augmented by local rain, becomes thicker and its flow again becomes turbulent and therefore erosive. Such flow is, according to King, the agency essentially responsible for pedimentation, but only in so far as the water is not collected into streams (permanently?), for if it is thus concentrated gullying and dissection will take place. Very heavy showers (of more than 2 inches an hour) will give rise to sheetfloods, though King has not himself seen these; flow of the sheet must in such a case be turbulent, and from the start it must be capable of erosion. Light rains, on the other hand, produce no flowing sheet at all; the hillside furrows then extend across the pediment and tend to dissect it (though sheet wash will repair this damage at another time).

To sum up, various processes—rill work, sheet wash, sheetfloods, and under certain conditions of relief and structure lateral planation—can take part in developing the same pediment, for temporary dominance of one or another depends on the highly variable activity of the runoff. According to climate, kind of rocks, and stage of development also the process of planation will be hindered to some extent by the tendency of running water to corrade vertically, but once the process is established it gains ever stronger control, except in so far as it is affected by external interruptions.

Pedimentation operates on every scale. Though badlands are justly cited as an excellent example of pluvial dissection, Smith (1951) and Baker (1951) show that they may become miniature reproductions of mountain ranges partly consumed by pedimentation and having uniformly steep slopes that are separated by a re-entering angle from flat gully floors, and also developing embayments and cols that are lowered by the intersection of ravines opposed head to head. Water flows off partly in streams and partly as sheet flow. The only conditions necessary for such development are, on the one hand, sparseness of vegetation associated with the occurrence of occasional heavy showers and, on the other, rock that is easily eroded but is coherent, does not flow,

[16] The relevant passage has been reviewed, and the figure from it reproduced, by Cotton (1942, p. 48–49).

[17] King (1953, p. 735) mentions an interesting test distinguishing between laminar and turbulent flow: if the surface of the water is depressed where it encounters an obstacle, the flow is laminar, and if it piles up the flow is turbulent.

and furnishes uniformly fine debris free from coarse fragments impeding regular runoff.

As regards the question whether, as some maintain, the primary process is undercutting of the hillslope at the base and consequent extension of the pediment or whether, according to other views, spontaneous retreat of the hillslope results in an embryonic pediment trimmed thereafter into shape by the distributed or diffused flow of water (King, 1953, p. 730–732), it would appear that one or the other process is dominant according to the nature of the rock and the climatic regime. In any case they are interrelated. On the one hand, the hillslope can continue to retreat only if the debris derived from it is removed by wash, at least from time to time, from the foot of the slope. The more swollen the flood of runoff, on the other hand, and the more it is accelerated en route, the more promptly will it spread out as a sheet at the foot of the hillslope. Thus the hydraulic discontinuity will be better defined, other things being equal, the higher and steeper the hillslope and the more nearly horizontal the flat at its base. The process, whether initiated tectonically (by a fault or flexure) or simply by river entrenchment, will be intensified both as regards backwearing of the hillslope and downwearing of the pediment as long as the former is increasing in height while it is still retreating. It will dwindle again in height, but the load of solid debris will then have become smaller and the slope of the pediment will be gentler; yet there will still be a re-entering angle between hillslope and pediment; this is at least as conspicuous around minor nubbins as at the foot of the most imposing line of bluffs.

In conclusion it must be admitted, in agreement with King (1953, p. 733), that "the pediment . . . is the ideal form to dispose of the maximum volume of water in minimum time and with least erosional damage to the landscape." In other words, like every form of mobile equilibrium (Baulig, 1950, p. 41) it is a form "of least work". It is indeed a form that is "stable" (King) in the sense that it retains its shape while it is being re-graded (though increasingly more slowly) throughout the cycle. It does not follow from this, as King claims, that it is the only or almost the only form of subaerial planation.

Pediplains or Peneplains?

According to King a pediplain and a peneplain must be immediately distinguishable by the predominance of concave surfaces on the former and convex on the latter. Certain passages in Davis seem to justify this interpretation. Thus, Davis (1912, p. 65–66, Fig. 25) has sketched the evolution of a mountain mass dissected under the vigorous attack of "normal" erosion. As long as valleys are being deepened crests can be sharp (though their sharpness depends on texture of dissection and valley depth); but when incision of valleys slackens the hillslopes are lowered, the crests become rounded, and, as relief becomes smaller, the ridge-top convexities become broader while their curvature becomes less sharp. (Lower slopes are not here in question.) Later Davis (1932, p. 409) added to this a statement that in a temperate climate "the concave basal curve . . . is enlarged as it retreats" (meaning rather "as the upper part of the slope retreats"). The need for this correction is shown by the accompanying figure (p. 409; reproduced by King, 1953, p. 723), which shows at an advanced stage a concave basal curve about three quarters of the length of the whole hillslope.

Accurately drawn profiles show, moreover, that on fully mature slopes, even on a permeable rock like chalk, the concave part of the slope is longer than the convex part (Baulig, 1940; reprinted 1950, Fig. 1, p. 129, profile A: local relief 20–30 m) and that on argillaceous sands it is still longer (profile C: local relief 30 m). The very carefully measured profiles of Fair (1947) from the plateaus of the interior of Natal confirm this. Alternating sandstones and thin-bedded shales provide a practically homogeneous terrain only moderately resistant, but this is capped by very resistant dolerite. As long as the dolerite cap survives the profile remains for the most part concave, for the slope is determined to a great extent by the veneer of dolerite blocks derived from the cap (becoming smaller down the slope). When the hard-rock cap has gone the profile becomes convexo-concave, and the summit convexity extends farther and farther, though even at the most advanced stage shown by Fair (with maximum declivity 6°) it occupies only 800 feet of the total length, 2800 feet, of the hill-

slope. The interior of Natal has a semi-arid climate (rainfall 25–35 inches a year, but strictly seasonal), but the passage from the convex to the concave segment is quite gradual (though Fair describes the concave slope as a pediment).

Pallister (1954; 1955) has shown that profiles like those described by Fair in Natal are found almost on the Equator, though in an equable and moderately humid climate in South-central Uganda (annual rainfall over 45 inches; savanna vegetation: elephant grass and a few large trees; complex crystalline metamorphic bedrock structure, scarcely affecting the relief; soils everywhere thick, in many places over 50 feet, over weathered rocks). The land surface is moderately undulating to low and hilly (relief up to 500 feet). The highest hills have flat tops capped by relics of a duricrust of laterite or a ferruginous material (ironstone), indicators evidently of a phase of semi-arid planation dating from mid-Tertiary times. The profiles are thus strongly concave, their steepness decreasing rapidly from 25° at the top, below the edge of the carapace, to 7° (average) on most of the slope. In contrast with this topographic form lower hills—less than 200 feet high—which evidently represent a more advanced cyclic stage have gentle and regular profiles; a convex summit segment of moderate length passes imperceptibly into a long concave slope inclined at 5° to 7°. Incision of the stream courses to a shallow depth at the foot of the concave slopes may be ascribed to a small recent climatic change, but the streams are sluggish, their floors are swampy for the most part, and their solid load consists of fine silt only. Is not such relief approximately that of the Davisian peneplain stage?

There is justification, therefore, for believing that, allowing for rock and climate, a continuous gradation exists from the gently undulating peneplain with monadnocks to the almost ideal pediplain surmounted only by rock stacks. The great difference between these two is that pedimentation produces at an early stage some planed surfaces which afterwards extend gradually, coalesce, and replace one another; the result is the production of pediplains of vast extent. Peneplanation results after there has been an increasingly slower reduction of relief over the whole area affected, and it is controlled throughout a cycle by a single base level. Furthermore, peneplanation is to a much greater extent than pediplanation under the control of lithological variations, and consequently does not produce surfaces comparable in areal extent and perfection of development to pediplains, at least on terrains resistant to humid-climate processes. Not all planed land surfaces, however, are either subarid pediplains or "atrophied" or "degenerate" pediplains; though this is what King sets out to demonstrate.

It must be admitted that he is right concerning not only the origin of South African, Australian, and other analogous planed surfaces, but also, in general, planation on resistant rocks in times anterior to the middle of the Tertiary (Baulig, 1950, p. 34; 1952a, p. 256–258; 1952b, p. 224).[18] The reason for this is obvious: as the higher forms of vegetable life on the land date only from the middle of the Paleozoic, rocks were until then exposed without protection to the direct action of meteoric agencies—wind, rain, eventually ice—whether the climate was humid or arid. The deserts of those times were biological, not necessarily climatic deserts. Woody plants existed from the Devonian onward; yet without dense undergrowth or thick litter (leafy trees date only from the Tertiary) the soil could have little protection from rain wash. Grasses in particular, such as form continuous turf, did not appear, it would seem, or at any rate were not widely distributed, until the middle of the Tertiary; only then were the biological conditions favoring peneplanation fully realized in humid-temperate climates.

On the other hand, sedimentary deposits and residual soils afford evidence of a number of periods before that when more or less arid climates prevailed over much of the earth's surface—in the Devonian, Permo-Triassic, and early Tertiary. The surfaces developed at such times are presumably pediplains. Yet aridity prevailed only in the intervals between humid periods, which are evidenced by coals and lignites formed from vegetation that accumulated in bodies of fresh water and thus indicate the existence of perennially flowing rivers. Such

[18] Barrell (1916, p. 361) was well aware of the importance of vegetal cover for sedimentation and consequently for continental denudation.

deposits, and more particularly the climates that favored their formation, may have been very widespread, for relics of them have survived only owing to special circumstances, namely immediate burial of the vegetal debris under detrital materials, which preserved it from oxidation, combined with persistent subsidence, which has afforded protection from erosion in later times. Planations that were developed in these humid periods are almost unknown, however, perhaps because humid-climate soils are easily removed and replaced by others that are very different, whereas soils developed under arid and subarid climates have a much better chance of preservation (*supra*).

Whatever the reason, the later Tertiary was destined to be the age of peneplains. But it is a fact that scarcely any of the known planed surfaces dating from this period have great extension over really resistant rocks, for example the Central Massif of France. This should perhaps be ascribed to the shortness of the period, to changes of base level that caused a succession of cycles of erosion, or, more especially, to slowness not of dissection but of the eventual planation in a temperate climate. In middle latitudes there are, however, almost everywhere locally planed surfaces that bevel moderately resistant terrains even as young as Pliocene. Some authors, King in particular, have preferred to regard these as ancient subarid pediments now to some extent degraded and disguised by humid erosion. The late Tertiary floras of western Europe do not confirm this; they indicate a climate significantly hot and humid, comparable to that of North and South Carolina and Louisiana, though it became colder at the approach of the Pleistocene; but they record no arid or semi-arid episodes. Pedological indications, though sometimes utilized, are unreliable: red soils are forming today in the Mediterranean region and on the Atlantic Piedmont of the United States, but these have quite different climates and contrast strongly in the seasonal distribution of rain.

The most significant fact bearing on this question is probably the occurrence of the "rañas", those vast mudflow deposits that were spread out in the late Pliocene over well-leveled surfaces in the western part of the Spanish Meseta and in southern Portugal (Hernández-Pacheco, 1950; Ribeiro and Feio, 1950). In these deposits angular blocks of all sizes, but consisting only of rocks particularly resistant to chemical weathering, are mixed with argillaceous-sandy materials that have resulted from advanced decay of rocks of all kinds. This debris furnishes evidence of profound chemical alteration of the terrain under a hot and humid climate. (That such a climate prevailed in those regions until rather recently is proved by the survival in the Pliocene of southern Portugal of certain tropical plants, ferns and others, that had disappeared by that time from Central Europe.) Stripping off of the debris that had thus been prepared by weathering, though facilitated by some tectonic movements, must have been due to a drastic thinning of the vegetal cover, but the cause of this was not necessarily pronounced (or prolonged) aridity like that of the Atlantic Sahara, with which it has been compared. The areas that were affected are indeed now the driest in Europe because of their low latitude and their situation with respect to the ocean.[19] Aridity, though it was perhaps not much more pronounced in the late Pliocene than at present, was sufficient to upset a delicate equilibrium between vegetation and soils that had become established in the preceding long humid period. Whatever the reason for this it would not be justifiable to extend to other regions conclusions valid for parts of the Iberian Peninsula. In the Appalachians or the uplands of Belgium, for example, there are indications of a former climate different from the present one, but this is rather a climate characterized by cryergic[20] activity, otherwise called "periglacial".

It is becoming more obvious that such cryergic activity has affected a large part of the lands now under a temperate climate. Considering the shortness of postglacial time it is legitimate to ask, therefore, whether most "humid-temperate" landscapes may not be periglacial landscapes merely retouched by "normal" erosion. This is a line of thought

[19] The authority for this statement is M. Sorre's aridity map in de la Blache and Gallois (1934, p. 81).

[20] Translator's note: Professor Baulig has recommended that "cryergic" be used instead of "periglacial" in the sense in which the latter word is commonly but "rather incorrectly" applied to processes in some regions not actually peripheral to glaciated regions (Baulig, 1956, paragraph 77).

followed by King (1953, p. 735) and by Cotton (1955, p. 1028). The latter author, recording changes effected by solifluxion in the middle district of New Zealand, at Lat. 41° S., writes: "The preglacial form of the surface [an upland peneplain; altitude about 1100 feet] must almost certainly be now changed beyond recognition or restoration." The modification that has certainly taken place has been variable with the nature of the rock, the strength of the relief, and the extent and duration of the cryergic activity. Further, we see today an end condition, the cumulative result of climatic changes that have succeeded one another through and since the Pliocene. Each phase, whether glacial or interglacial, perhaps even interstadial, has set about the task of remodeling to its own norms the forms bequeathed to it from the preceding phase. It appears generally, however, from the modification of relief forms and from the extent to which soils are weathered, that some or all of the interglacial have been longer than the glacial ages. Because of repeated displacements of the climatic zones in the Pleistocene it is difficult to find landscapes that have always been under humid-temperate conditions of climate, and this poses very serious problems. We can, nevertheless, attempt to reply to the following questions: Have major forms of the landscape as well as detail forms been remodeled by cryergic processes? and, in particular, have the profiles of mature hillslopes been profoundly modified?

In the cryergic regime disintegration and comminution of materials are essentially the result of repeated freezings—of gelifraction. Ablation of debris is effected by free fall, by a kind of dry creep, and by runoff of rain or snow-thaw water; but the most potent agency is solifluxion[21] (or "gelifluxion"). This process is slow displacement of water-saturated thawed material which slides over a still frozen subsoil (not necessarily perennially frozen ground) the impermeability of which prevents drying out that would bring the fluent mass to a halt. It is a true mass movement, resembling mudflow; but, whereas mudflow is sporadic and is usually chaneled into a definite course, gelifluxion sprawls over all slopes that are neither too steep nor too gentle for it. Further, the diurnal thaw takes place practically simultaneously throughout the extent of the debris sheet to a depth that fluctuates throughout the warmer season but varies little from day to day. The mass movement may be thought of as quasi-continuous and relatively uniform throughout the season in which it can take place; and this is a condition strongly favoring development of a smooth profile. Tundra vegetation checks movement, it is true, and as a result garlands and terracettes are formed; but these minor accidents do not significantly change the fundamental nature of the phenomenon.

Movement of this kind, massive and sprawling as it is, tends to fill up hollows in its path; and the first to be filled in are valley bottoms, in which considerable accumulation has commonly taken place. It is not necessary to assume on this account, as some have done, that there has been scarcity of water streams in glacial ages; rather has there been considerable augmentation of the solid load carried by streams; for valley aggradation implies merely a tendency towards an equilbrium better adapted to changed conditions. In this way a V-shaped ravine can become filled with "head" to the extent that the stream of water that originally cut it is pushed to one side, where it excavates a new trench, whereas the tongue of soliflual debris, now dried out and therefore stagnant, remains convex. There can thus be local inversion of relief (Cotton and Te Punga, 1955b).[22]

In contrast with this, gelifluxion is known to corrade weak rocks to some extent. Recently various authors (King, 1953, p. 736; Te Punga, *in* Cotton and Te Punga, 1955a, p. 1008) have observed that where the rock is not very resistant the layer of soliflual debris commonly rests on a surface so remarkably smooth that it might have been shaved with a plane. Cryergic corrasion has even produced some concave slopes. It was long ago observed that valley heads that are feebly incised in an upraised peneplain commonly have a gently con-

[21] We could perhaps call it more correctly "geli-(soli)fluxion". It seems unjustifiable to restrict the meaning of solifluxion (Andersson, 1906) to cryergic phenomena. Flow of soil may be seen in all climates, and it plays a first-order role on slopes in the super-humid tropics.

[22] This process is analogous to that described by Kirk Bryan (1940) in material with a rather special structure.

cave, or catenary, transverse profile, having an appearance of late maturity. This was supposed by Davis to be due to the feebleness of infantile streams, which have failed to cut deeper (but must be judged to have been capable nevertheless of evacuating the debris resulting from degradation of the side slopes?). Cyclic analysis (Baulig, 1935, p. 15, Fig. 5; Sittig, 1936) seems, on the other hand, to indicate that the last traces of ancient mature valleys survive in such cases; the downstream parts of these can be restored by extending transverse profiles preserved on spurs and also the longitudinal profiles of tributaries; and accordance of profiles thus plotted appears to be satisfactory. It has later been recognized by Macar (1954, p. 20–21), however, that such small valleys high up on the Ardennes are encumbered with soliflual debris, though it is not known whether this is a thick layer even on valley floors, and it does not seem to be thick enough on lateral spurs to invalidate the reconstructions that have been made of profiles. Cotton (*in* Cotton and Te Punga, 1955a, p. 1019 *ff.*) has further suggested that these "dells" (equivalent to the German *Dellen*) may be of periglacial origin. They are not, he says, nivation cirques, and neither are they glacial cirques, because their smoothly concave profile excludes an origin by basal sapping of slopes; and, moreover, true cirques are unknown in the region described except perhaps at much greater altitudes. He is of the opinion, therefore, that a "normal" V-shaped section has been widened out and deepened by channeled solifluxion. He has not been able to ascertain, however, whether there may be some considerable depth of geliflual debris in the valley bottom.

It would appear then that gelifluxion has plastered over minor landforms and that accumulation of permeable "head" must have obliterated the smaller ramifications of stream systems so that texture of dissection has become coarser (Cotton, *in* Cotton and Te Punga 1955a, p. 1016).[23] It may even be that in incoherent materials, ground-morainic debris for example, cryergic processes have been able to obliterate glacial relief altogether (it would be weak in any case) and to substitute for it relief of their own making; but are we justified in assuming that they have reduced or softened to any appreciable extent more rugged forms on rather resistant rocks? In particular, have they broadened the summit convexities above mature hillslopes (Cotton, *in* Cotton and Te Punga; 1955a, p. 1028 and Fig. 6)? Rather they must have lengthened concave slopes at the expense of summits and thus reduced the radius of summit convexity.

A hillslope affected by gelifluxion is really a "covered slope" throughout its length, and, as the mobile regolith will thus impose its own profile, the underlying rock will the more readily yield debris of increasingly fine texture, and this, being saturated with water, will be increasingly mobile. This holds for the deeply weathered bedrock greywacke of the locality described by Cotton. Gelifluxion and the corrasion due to gelifluxion produce little change on a broadly convex summit, but down the slope, with abundant water, with increasing comminution of debris, and with increasing thickness of the layer of debris, mobility increases; and thus an ever greater quantity can be evacuated down a diminishing declivity, so that concavity of profile develops. Further, movement starts not far from the summit, probably higher up on the hillslope than where rain wash becomes concentrated on a "normal" slope and produces concavity; and the movement continues towards the foot even on declivities of a very few degrees—apparently lower declivities than those of "normal" slopes. The profile will be flattened, therefore, in its lower part and steepened in its upper part, eventually to the extent of exposing a rock face (Peltier, 1950); thus its concavity is increased, whereas the convex portion, reduced in breadth, becomes more sharply curved ("rounding of summit convexities", Cotton, *in* Cotton and Te Punga, 1955a, p. 1028). It is here assumed that the relief was originally rather vigorous; a more gently undulating surface will, on the other hand, be further flattened by lowering of the higher parts and building up of the hollows.

Conditions that favor extension of the concave segment of the profile are the same in a

[23] Davis believed that the same thing was to be expected at an advanced stage of the "normal" cycle as a result of increase of thickness of the permeable mantle of residual debris.

temperate and in a cryergic regime. A mild, humid climate assures an abundant runoff, whereas in the cryergic there is frequent alternation of nightly frost and daily thaw during a long season that is relatively warm. If in either climate the rocks are already shattered their disintegration and the comminution of the debris is facilitated, and if argillaceous materials are abundant both runoff and mobility of the regolith are thereby facilitated. Vegetation will be more strongly inhibitive of downslope movement in the temperate climate, but *per contra* it engenders biochemical action that produces colloids, and such action is practically negligible in a cryergic regime. Gelifluxion appears, on the other hand, to be a more powerful agent of degradation on slopes than are humid-temperate processes. It does not seem, however, that on the same rocks gelifluxion produces profiles that are radically different; and thus it is difficult to evaluate the part, probably varying greatly from place to place, it has played in shaping the present landscape. No doubt it would be possible to find a reasonably good solution of this problem by observing the attenuation of cryergic activity from higher to lower latitudes, but comparisons must be made on land surfaces underlain by similar rocks.

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4 Rue Fischart, Strasbourg, France

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Short Notes

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PALEOMAGNETIC SURVEY OF TRIASSIC ROCKS FROM ARIZONA

BY PAUL R. KINTZINGER

During the summer of 1955 directions of magnetization were determined for rock samples from Triassic formations exposed along the Echo Cliffs and Vermillion Cliffs, near Marble Canyon, Arizona. Collections were made from fine-grained red sandstones and siltstones of the Moenkopi formation and the Dinosaur Canyon member of the Moenave formation. The Moenkopi formation in this area is about 230 feet thick. Samples were collected throughout

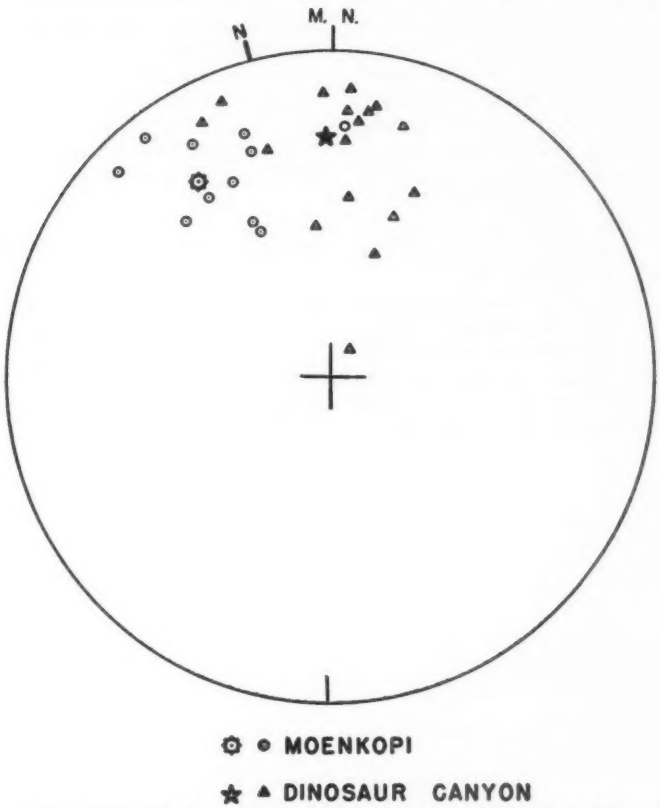

FIGURE 1.—STEREOGRAPHIC PROJECTION OF DIRECTIONS OF MAGNETIZATION OF TRIASSIC ROCKS FROM ARIZONA

TABLE 1.—DIRECTIONS OF MAGNETIZATION OF TRIASSIC ROCKS FROM ARIZONA AND NEW MEXICO

Formation	Location of Samples		Declination	Inclination	Investigator
	Latitude North	Longitude West			
Springdale	37°15′	113°0′	N. 9.75° W.	+38.8°	Runcorn
Dinosaur Canyon	36°51′	111°34′	N. 21° E.	+30°	Kintzinger
Chinle	35°23′	104°47′	N. 48° E.	+32°	Graham
Shinarump	35°53′	111°25′	N. 4° W.	+33°	Graham
Moenkopi	36°51′	111°34′	N. 35° W.	+35°	Kintzinger
Moenkopi			N. 0.2° E.	+27.4°	Runcorn

this thickness. The Dinosaur Canyon formation was sampled over a thickness of more than 80 feet. The lateral extent of the sampling of both formations was greater than 1 mile.

The directions of magnetization are shown in Figure 1. For all specimens the north-seeking end of the magnetization vector was directed downward. In Figure 1 the results for the Moenkopi are indicated by small circles, with the mean direction of magnetization indicated by the large circle. The Dinosaur Canyon results are indicated by small triangles, with the mean direction indicated by a star.

The mean directions of magnetization have been listed in Table 1 in stratigraphic sequence along with results obtained by Graham (1955) and Runcorn (1956). It is interesting that, although the values for the declination range from N. 35° W. to N. 48° E., the values for the inclination are relatively constant. Runcorn's (1956) results from the Moenkopi and Springdale sandstone were taken from pages 312 and 314 of his publication.

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NEW MEXICO INSTITUTE OF MINING AND TECHNOLOGY, SOCORRO, NEW MEXICO

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